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Especially when using older guidebooks in this collection, note that locations may have changed drastically. Likewise, geological interpretations may differ from current understandings.

Please respect any trip stops designated as “no hammers”, “no collecting” or the like.

Consider possible hazards and use appropriate caution and safety equipment.

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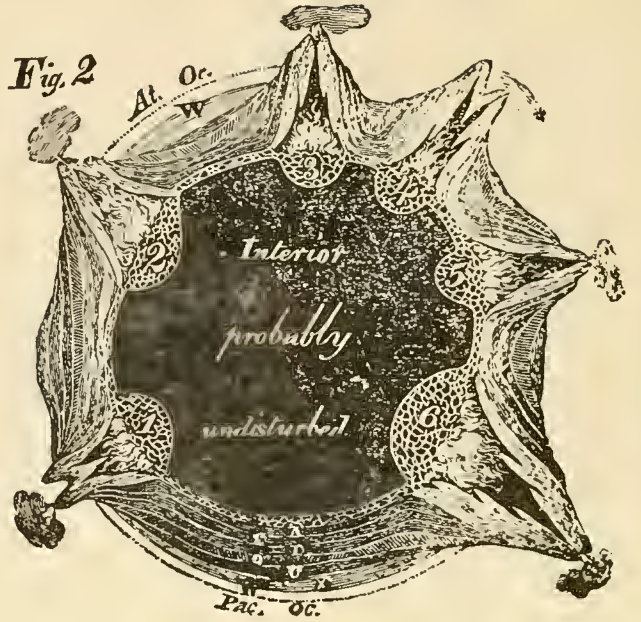
New England Intercollegiate Geological Conference 1969

Fig. 1.

Fig. 2



SEGMENTS.

EXPLANATIONS.

Formations.

Figures 1 and 2.* A. Lower carboniferous formation—E. Lower quartzose formation—I. Lower calcareous formation—O. The lower side of the second carboniferous formation when used in Fig. 1. This lower part is probably primitive. When used in Fig. 2, it is for all that formation—U. Second quartzose formation—X. Second calcareous formation—W. Oceanic waters.

Combustibles.

These are indicated by numeral figures. In Fig. 1, they are represented as they are supposed to have been deposited at the creation. In Fig. 2, they are represented as having been consumed by combustion, whereby an explosion was produced, which burst through the primitive and transition series—the only deposites then made; and those not perfectly indurated. 1. The combustibles under Rocky Mt.—2. New-England—3. Britain—4. Alps and Pyrennes—5. Caucasus—6. Himalay.

Remarks. In Fig. 1, the water is represented as encompassing the whole earth; being pressed out to the surface by the greater specific gravity of the earthy materials. While the earth and waters were in this quiescent state, no organized beings, but marine, were provided with a place of residence. In due time the combustible materials marked 1, 2, 3, 4, 5, 6, were ignited, and produced the changes exhibited in Fig. 2.

* These figures are an improvement upon those published in my Geological Index, in 1820, and afterwards copied into Woodbridge's Geography.

NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

61st Annual Meeting

at the

State University of New York at Albany
Albany, New York
October 10, 11, 12, 1969

GUIDEBOOK FOR FIELD TRIPS IN NEW YORK, MASSACHUSETTS, AND VERMONT

Editor

John M. Bird

State University of New York at Albany

1969

NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

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Guidebook for Field Trips in New York,
Massachusetts, Vermont

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EDITOR'S PREFACE

Attendance at N.E.I.G.C. meetings has "exploded" during the past decade. This year over 700 people have pre-registered, so that it is essential that enough field trips are available to serve all the varied interests such attendance represents. This Guidebook has been prepared by the field trip leaders. Because of its size, I have done no more than assemble the articles into this volume; the authors have done their own editing.

The State University of New York at Albany has subsidized the cost of printing this Guidebook. Additional copies will be available from the SUNY-A Bookstore, Albany, New York 12203.

The articles appear in the order used for registration announcements. Many of this year's field trips go considerably west of geographical New England. However, I think all will agree that all of the trips are within the confines of geological New England and, therefore, of interest to the N.E.I.G.C.

I record here our thanks for the significant contributions and generous efforts of the field trip leaders. Also, we thank Dr. Peter C. Benedict and Miss Linda Schroll for their considerable effort on the logistics for this year's meeting. E-an Zen provided the photo for the inside front cover and, thereby, a good reason for me to include a bit of wisdom and amusement provided by one of the first geologists to study the region encompassed by this year's meeting. It is from:

Eaton, Amos, 1830, Geological Textbook, prepared for popular lectures on North American geology; with applications to agriculture and the arts: Albany, Websters and Skinners, 64 p. (figs. 1 and 2 are on p. 18-19; on p. 17 is the short caption, which reads: "Exhibition of two transverse Segments. The earth is here supposed to be cut into two parts, at the 42° of north latitude. The observer is supposed to stand south of the center of the segments -- all the earth, south of him, being removed.")

PREFACE.

Every geologist is, probably, more or less misled by theory. If the earth was washed and the rocks left clean, they would not disagree in regard to rocks. But they are now dependent on naked cliffs and deep river-washed ravines; which present to the eye less than a hundreth of the evidence required. For ninety-nine hundredths, theory alone furnishes the facts upon which the very <u>same theory</u> is founded. But a long course of observations and careful comparisons, have done considerable towards a correct system of generalization.

Geology is subject to an evil peculiar to itself. If its votaries disagree, the common learner has neither time nor inclination to review their data, by visiting localities referred to, and thus to correct their mistakes and adjust their differences. The reasons, fully written out, which govern the experienced geologist, would require numerous octavos for each stratum. Therefore the learner must rely upon his confidence in his teacher's habits of careful investigation, his fidelity, his independence, and his talent at generalization. He must not overlook the advantages presented by the district of country which he examines. He should therefore compare the districts examined by different geologists.

The little island of Britain can furnish no authority for a general system; though the industry of distinguished geologists has done much towards an elucidation of important points, in detail. France, taking in the Alps as a primitive nucleus, presents more advantages. But according to De Luc, America alone must give a system of general strata. How far I have succeeded in my attempt to present such a system, future investigations (not the opinions and closet speculations of the geologists of either continent) must decide. For the system adopted in this text-book, I rely on my own personal examinations, aided by Dr. T. R. and Prof. L. C. Beck, M. H. Webster, and J. Eights, more or less supported by Professors Hitchcock, Dewey, and Emmons, from the Atlantic to the western extremity of Lake Erie. For the remainder, I rely upon the personal examinations and collected specimens, which I have now before me, of Dr. Zina Pitcher and Dr. Edwin James. Messrs. Schoolcraft and Peter have also contributed much. We have, altogether, traversed a succession of northerly and southerly strata through more than forty degrees of longitude.

A text-book is too small a name for these days of puffing arrogance. But I propose to present all my supposed heresies to the geological fraternity in this form and under this title. And I beg the favor of the most regorous criticism upon this book, small as it is. To stimulate men of science to the work of examination and of criticism, I will state; that I intend to publish considerable in scientific journals, also a full system, upon this plan. As I have had more than seven thousand pupils already,* and shall probably have more still, it will be well for them "to be on the alert" if I am propagating errors. I am not in sport--I have, during the last fifteen years, travelled over seventeen thousand miles, for the express purpose of collecting geological materials; the results of which are comprised in this little octavo pamphlet, and exhibited in the accompanying map and wood cuts.

I may be accused of fickleness on account of the changes which appear in every successive book I publish. I confess this is the ninth time I have published a geological nomenclature; and that I made changes in each of more or less importance. But I have always consulted my scientific friends; and every change was founded on new discoveries in "matters of fact." In this text-book, the principal changes relate to the graywackes. The Allegany mountains, I had never examined before with particular care. I verily think, these mountains present everything required for settling that part of the science. The various deposites of Detritus had not been thoroughly studied by any American, when I published my last nomenclature. I believe I have made a few changes in that department, which will finally obtain. I now adopt the Tertiary formation of Europeans; but I find no facts here to justify their numerous subdivisions.

*Rather, auditors.

I can now give European equivalents for all our strata, excepting the ferriferous and geodiferous. Our fourth series, however, seems to be more solid, harsh, and vastly more extensive, than its supposed equivalent. Perhaps it is a repetition of our third series. It is certainly distinct in the range of the profile given at the foot of the map, however.

With all deference to the high character of De La Beche; as an experienced teacher I may say, that his numerous sub-divisions, if adopted, will ruin the science.* Others have done much towards driving the study from our schools, by introducing petty local names. If their authors were not entitled to high respect, on account of other services to the science, one would feel disposed to treat such names ludicrously. For example, there is a variety of first graywacke in a place called Pilfershire, in Columbia county, remarkable for enduring heat. There is another in a place nicknamed Fuddletown, in Onondaga county, of a cellular texture, much used at Salina. The former should be called Pilfershire stone, and the latter Fuddletown stone, to be equivalent, in absurdity, to Purbeck, Bagshot, and other ridiculous European names.

The distribution of strata into five series, cannot be called an innovation; for it produces no change whatever. It amounts to nothing more, than referring well established strata one step further back, towards an elementary basis.

Students, for whom his text-book is intended, may feel no interest in any thing personal, relating to myself. But I will throw this paragraph in their way. I have been accused of arrogance for stating facts relating to American geology, without formally bowing to European authorities. I should condemn myself for any step in the science, which was not taken with a due consideration of all that had been done in Europe, Asia and Africa, in advance of our own investigations in point of time. Whoever is "first in the field" of natural science, has an exclusive right to give names. His successors should either adopt his names, or give them as synonyms and equivalents. This is essential to the very being of science. But English and French geologists have introduced new names, not adopted in Germany; because new discoveries made them necessary. I have done the same thing in America, and for the same reasons.

I confess that this is a kind of "ipse dixit" text-book. It is so, because the plan does not admit of demonstration. In a future publication, I intend to cite authorities from nature to illustrate my views. But I am prepared to abandon any of them; as I have frequently done heretofore, in cases of numerous errors, to which I am still subject.

Geology is a progressive science; and he, who has any respect for his future reputation, should be exceedingly cautious about committing himself on matters of fact or speculation. I confess, that I have, most egregiously, violated this rule; but there are peculiar circumstances in my case, arising from my being "a hireling drudge" to the most munificent patron of this science, which will palliate, at least if not justify.

*See Table of Equivalents at the end of this Text-Book.

I despise arrogance; but I am within sixteen years of the "three score and ten," when the mind of man is averaged beyond the period of vigorous effort. About two score of these years have been devoted to Natural Science. I offer this as an apology for some dogmas, forbidden to youth.

AMOS EATON.

Rensselaer School, Troy, N. Y.
January 23, 1830.

Trip 1

STRUCTURAL AND STRATIGRAPHIC RELATIONS ALONG THE PRECAMBRIAN FRONT IN SOUTHWESTERN MASSACHUSETTS*

by

Nicholas M. Ratcliffe
City College of C. U. N. Y.

INTRODUCTION

Gneissic rocks of the Berkshire Highlands are at present undergoing the close scrutiny of several U. S. G. S. workers. I am indebted to my colleagues David Harwood, Stephen Norton, and Robert Schnabel for helpful discussions of stratigraphic and structural problems we share. My work began as a study under R. H. Jahns, at the Pennsylvania State University in 1962. Two generous grants from the Penrose Fund, and financial support from the U. S. G. S. in the Ashley Falls and Monterey quadrangles have allowed continuation of this project to the present.

The ideas presented in this field trip guide are more in the nature of a progress report than a finished product and are subject to changes as the mapping continues. Map data from the Great Barrington, Stockbridge, Ashley Falls, East Lee, and Monterey quadrangles have been used in this compilation (Ratcliffe, 1965, and unpublished data).

At the latitude of Stockbridge, Massachusetts, the Berkshire Highlands make an abrupt swing to the west in a large bulge of rock known as Beartown Mountain (Fig. 1). At this point the Precambrian front is offset approximately 5 miles to the west, and map distribution of the Cheshire Quartzite and Dalton Formation changes from a fairly regular belt to a highly irregular pattern of isolated outliers. This field trip will deal specifically with the stratigraphy and structure of both the Precambrian and Paleozoic rocks related to this bulge.

STATEMENT OF THE PROBLEM

The probable Lower Cambrian clastic rocks of the Cheshire Quartzite and Dalton Formation are present in thrust slices that rest above either the

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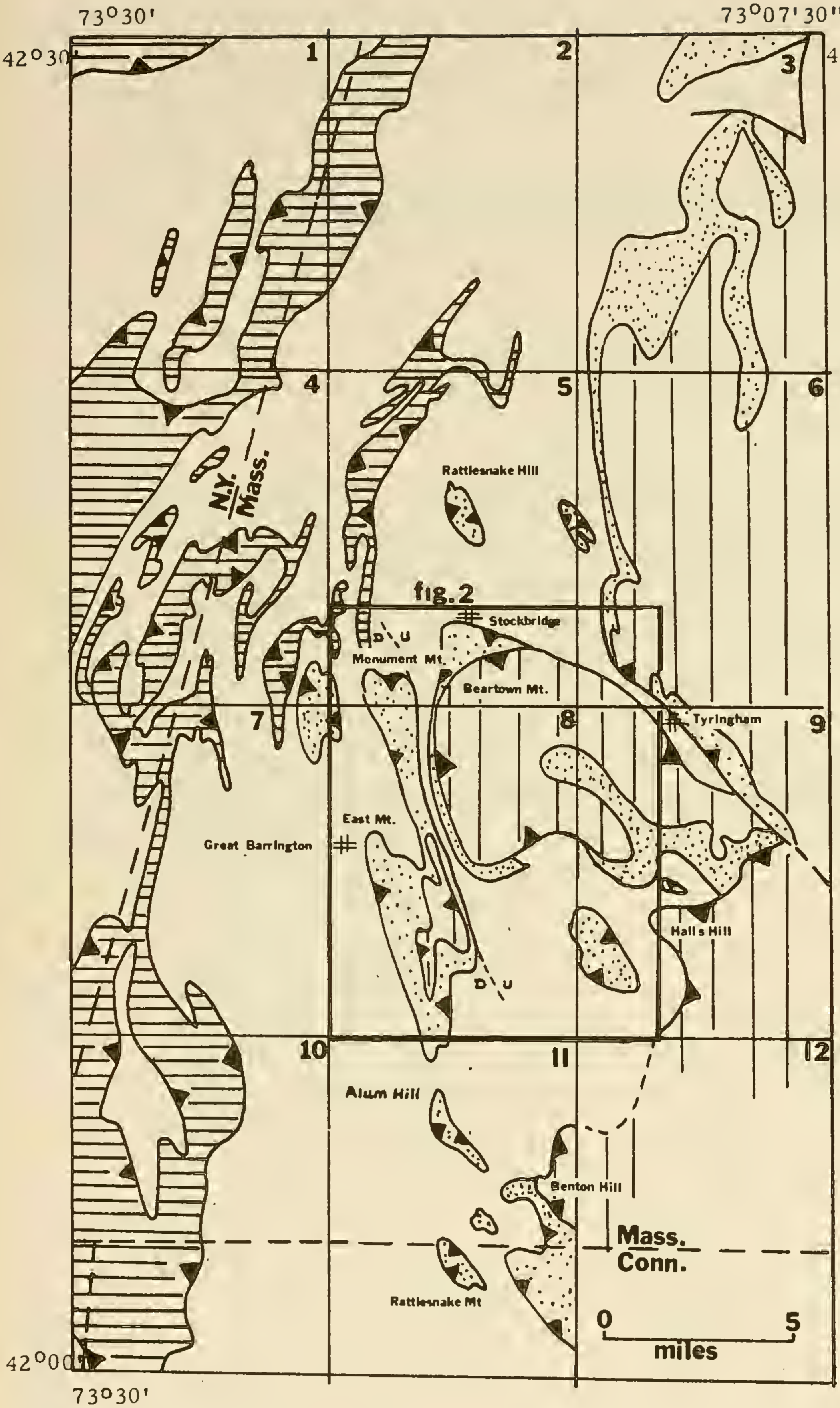
*Publication authorized by the Director, U. S. Geological Survey



Tectonic units

Allochthonous rocks of the Taconic sequence (Chatham and Everett slices).

Autochthonous rocks, comprising Walloomsac and Sto bridge Formations or local Cheshire and Dalton Forma and Precambrian rocks tha appear to be in place.

Parautochthonous rocks, co prising rootless exposures Cheshire and Dalton Forma and Precambrian gneisses. Stippled pattern - combined Cheshire and Dalton Forma vertical ruled pattern - Pre brian gneisses. Unstudied tacts in East Lee and Pittsf East quadrangles are shown without fault and may be in place.

Index to quadrangles

1. Canaan
2. Pittsfield West
3. Pittsfield East
4. State Line
5. Stockbridge
6. East Lee
7. Egremont
8. Great Barrington
9. Monterey
10. Bashbish Falls
11. Ashley Falls
12. South Sandisfield

Fig. 1. Major tectonic units in southwestern Massachusetts.

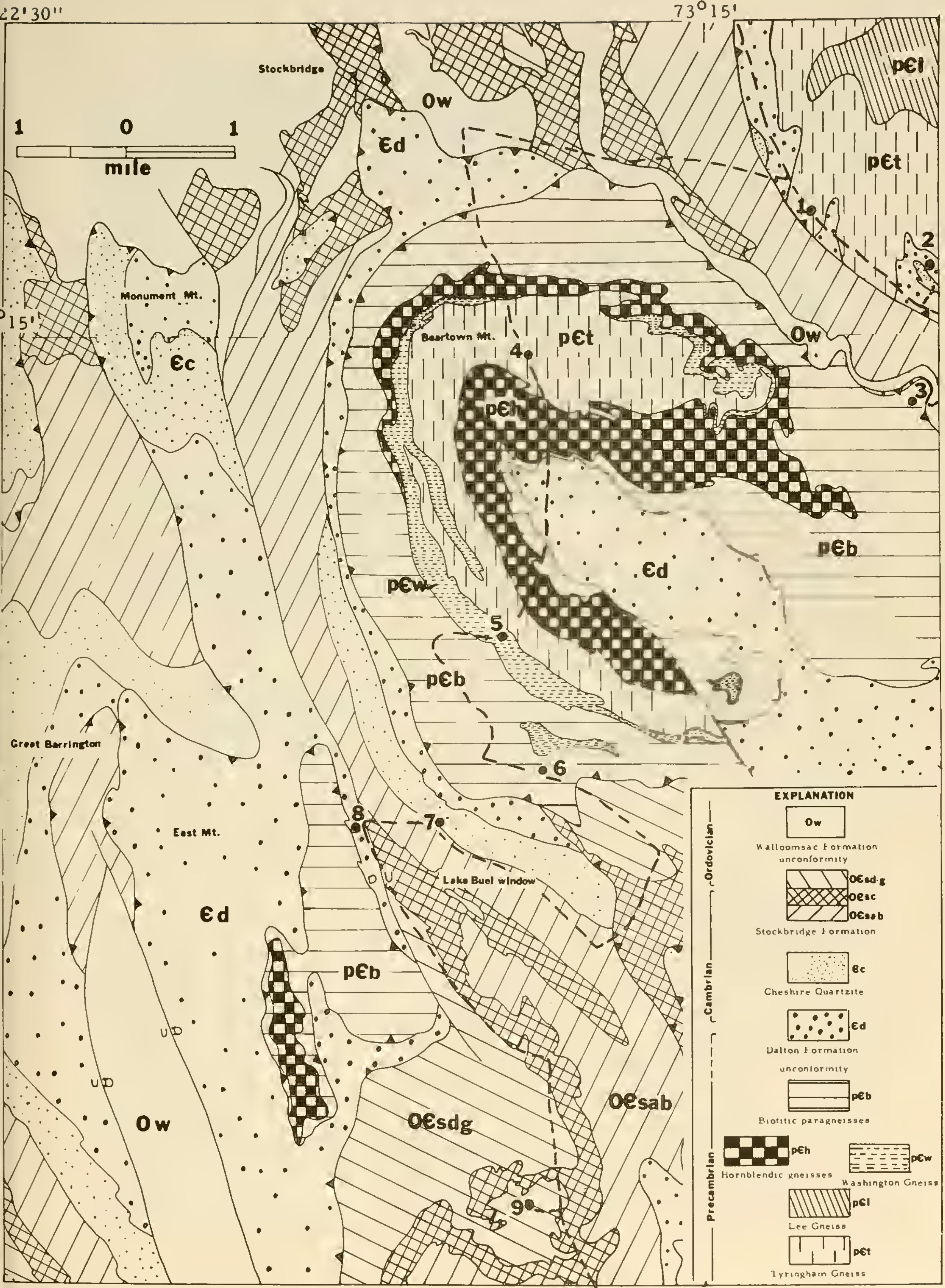


Fig. 2. Generalized geologic map of the Beartown Mountain area, showing location of stops.

Stockbridge or the Walloomsac Formation (Fig. 1). These rocks form an arcuate chain of outliers extending for 21 miles from Rattlesnake Hill southward to Rattlesnake Mountain. These outliers have been assigned to a single thrust system named the Monument Mountain slice for excellent exposures at that locality in the Stockbridge quadrangle (Ratcliffe, in prep.).

Precambrian rocks lie east of this belt and belong to a complexly folded and faulted sequence related to a large faulted nappe known as the Beartown Mountain slice. Inverted Paleozoic rocks flank this structure on the south, southwest, and west and everywhere dip under the Precambrian rocks (Stop 7). At the northeastern edge of the Beartown Mountain slice the fault passes into the air at Cobble Hill (Stop 3), where it is underlain by the right-side-up Stockbridge and Walloomsac Formations.

It appears likely that large parts of the Precambrian and Lower Cambrian clastic rocks in southwestern Massachusetts are not rooted and occur as large thrusts resting on the autochthonous Stockbridge or Walloomsac Formation (Ratcliffe, 1965, 1968). Assuming a thrust direction from the northeast to southwest, the minimum amount of horizontal transport of Precambrian rocks was seven miles. This horizontal offset probably resulted from the overthrusting of a large recumbent anticlinal nappe from the northeast.

To the north highly faulted and imbricated thrust structures are known to mark the Precambrian front in the Windsor quadrangle northeast of the Pittsfield East quadrangle (Norton, personal communication, 1969). The extent of thrusting of this kind along the western edge of the Berkshire Highlands is unknown because much of the area has not been mapped since the time of B. K. Emerson.

MAJOR TECTONIC UNITS

The major tectonic units recognized in the Great Barrington area listed from east to west are (1) a parautochthonous sequence, (2) an autochthonous sequence, and (3) an allochthonous sequence (see Fig. 1).

The parautochthon consists of Lower Cambrian clastic rocks and basement gneisses that have locally been overturned, thrust westward, and now occupy anomalous positions above rocks of the autochthonous sequence. The outliers of Rattlesnake Hill, Monument Mountain, East Mountain, Alum Hill, and Rattlesnake Mountain (Canaan, Conn.) constitute an arcuate band of the parautochthon west of the lobate bulge of Beartown Mountain. Beartown Mountain, partially rimmed by probable Lower Cambrian clastics (Dalton Formation and Cheshire Quartzite), contains a core of Precambrian gneisses and also is not rooted. An outlier of basement gneisses rests on top of the parautochthonous clastics at East Mountain.

The general synformal map pattern of Beartown Mountain requires a complex interpretation. The symmetrical map pattern (Fig. 2) indicates a single coherent fold. The Tyringham Gneiss occurs in the core of the Beartown Mountain structure, defining a synform that was rotated about a northwest-southeast axis by late folding, so that the flanks of the older structure are now nearly parallel.

The autochthon is composed of Lower Cambrian clastic rocks, the Stockbridge and Walloomsac Formations, and locally Precambrian basement rocks that appear to be overlain by right-side-up rocks. Within the autochthon a complex sequence of tectonic events is recorded involving four Paleozoic deformational phases, as summarized below.

D_0 - pre-Walloomsac: faulting, folding, and possible overturning. Angular unconformity preserved.

D_1 - post-Walloomsac: small isoclinal recumbent folds, possible slump folds related to Taconic thrusting, emplacement of Everett and Chatham slices.

D_2 - post-allochthon: regional northeast structural trends and development of major foliation (S_2) accompanied by low grade metamorphism.

D_3 - northwest-trending fold system developed, refolding all earlier folds and foliations. The intensity of this deformation increases eastward and controls the distribution of rock units in the Stockbridge, Great Barrington, and Ashley Falls quadrangles. The thermal peak of metamorphism developed during or shortly after this event. Thrusting of the parautochthon probably occurred at this time.

These features will be discussed at length in the field on Trip 2.

The allochthonous sequence (Fig. 1) occurs in the Chatham and Everett slices of the Taconic allochthon, consisting of the Nassau and Everett Formations of probable Late Precambrian or Early Cambrian age. Stratigraphic and structural relations in these rocks are discussed in Trip 2. A record of deformational phases D_1, D_2, and D_3 is preserved in the allochthonous Taconic rocks, indicating that thrusting of the Taconic sequence rocks took place early in the deformational history.

The deformation plan of the parautochthonous rocks differs markedly from that of the other two tectonic units. The Precambrian rocks were multiply deformed prior to deposition of the Lower Cambrian (?) Dalton Formation. These relations can be seen at Stops 1 and 2, where the

Precambrian-Dalton contact is nearly exposed. Paleozoic structural trends in the parautochthon are dominated by isoclinal or highly asymmetric shear folds having a strong axial plane foliation that is either subhorizontal or dips moderately steeply to the northeast. These folds commonly have a northeast over southwest rotation sense regardless of the attitude of the axial planes (gentle to the west, horizontal, or dipping moderately steeply to the east). A late set of open folds trending northwest-southeast locally arches these older features, but a foliation related to these folds is not apparent. The older folds appear to be related to thrusting of the parautochthon in a general southwesterly direction. Locally, rocks of the autochthon (Stop 10, northeast corner of Ashley Falls quadrangle) have this deformation plan as well.

The northwest-trending fold system in the autochthon and allochthon either predates or is synchronous with the peak of the metamorphism (Ratcliffe, 1965, 1968). Radiometric data (Zen and Hartshorn, 1966) indicate that the last phase of recrystallization may be as recent as Devonian. This suggests that the overthrusting of the parautochthon might be a Devonian event as well.

METAMORPHISM

Paleozoic metamorphism affected all rocks in this area. The grade increases on a regional scale toward the east-southeast. Isograds of biotite, garnet, and staurolite trend N. 15° to 20°E. approximately paralleling the New York-Massachusetts state line. Staurolite-bearing rocks are exposed near Lion's Head in the Bashbish Falls quadrangle (Zen and Hartshorn, 1966) and in the southwest corner of the Stockbridge quadrangle. The assemblage staurolite-kyanite-plagioclase-biotite-quartz is found in rocks of the Walloomsac Formation in the southwest corner of the Great Barrington quadrangle and in the western part of the Ashley Falls quadrangle. Sillimanite-bearing Walloomsac is exposed along the eastern edge of the Ashley Falls quadrangle and at Canaan Mountain in the south-central part of the same quadrangle.

The rocks shown in Figure 2 all lie on the high side of the staurolite isograd and on the low side of the sillimanite isograd. Stop 10, in the northeast corner of the Ashley Falls quadrangle probably is located close to the projected sillimanite isograd. The location of the sillimanite isograd is uncertain here owing to the lack of exposures of the schistose facies of the Walloomsac Formation that contains sillimanite at appropriate grades elsewhere.

Relict monoclinic pyroxene, largely altered to green hornblende, and strongly sheared perthitic feldspars present in some Precambrian rocks suggest that moderate to high-grade metamorphism took place in Precambrian time prior to deposition of the Dalton Formation. However, no relict mineral assemblages have been found that establish the grade of this Precambrian metamorphic event.

STRATIGRAPHY

Precambrian Rocks

Rocks of probable Precambrian age crop out at Tyringham (Fig. 1), in the core of the Beartown Mountain synform, and on East Mountain, Halls Hill, and Benton Hill. These gneisses are assigned a Precambrian age because they are overlain unconformably at many localities on a regional scale by feldspathic quartzites and conglomerates that locally contain Olenellus fragments (Walcott, 1888, p. 235-236). These unconformable relations were reported by Pumpelly and others (1894, p. 11, 100), by B. K. Emerson (1899, p. 39-40, 45), and more recently by Herz (1958, 1961) and Norton (personal communication, 1969) in the northern part of the Berkshire Highlands. Excellent exposures of the unconformity can be seen on the bed of Day Brook near Dalton, Massachusetts (Pittsfield East quadrangle) (Emerson, 1899) and at Stop 2, this trip. This unconformity is also developed in the South Sandisfield quadrangle (Dave Harwood, personal communication, 1969) and in the Ashley Falls quadrangle (Ratcliffe, unpublished data). Throughout much of the Great Barrington and Stockbridge quadrangles the Dalton-Precambrian contact is tectonically disturbed (overturned and faulted) so that unconformable relations are difficult to demonstrate.

B. K. Emerson's subdivisions of the Precambrian rocks of eastern Berkshire County (1899) were later modified in his compilation of the geologic map of Massachusetts (1917). Considerable confusion of terminology now exists because of discrepancies between Emerson's two maps, thus making difficult the correct usage of the classical formation names such as Becket Gneiss, Hinsdale Gneiss, and Lee Gneiss. Because of this uncertainty in the meaning of the classical formation names, either local names established by Emerson (1899) or informal lithologic names will be used for new map units or those of uncertain correlation.

The Precambrian rocks in the area of Figure 2 appear to belong to four major map units. Abundant lithic variation within these generalized units accounts for many members that cannot be shown at the scale of the map. The relative ages of the units are not known with certainty because of the lack of primary sedimentary structures for telling stratigraphic tops and the fact that the intense recumbent folding results in sections having opposite geometric tops. The preferred, but somewhat arbitrary, interpretation places the core rocks (the Tyringham Gneiss) of the Beartown Mountain structure at the base of the section.

Tyringham Gneiss

The Tyringham Gneiss (Emerson, 1898, p. 18; 1899, p. 34) was named for exposures in the Lee-Tyringham area (Stops 1 and 2) and specifically referred to the rocks exposed in the core of the Beartown Mountain structure (Emerson, 1899, p. 59) (Stop 4). The unit is mainly light-gray weathering, pinkish-gray, granitic to granodioritic biotite gneiss with distinctive quartz rodding produced by multiple obliquely intersecting cleavages. The potash feldspar, in individual crystals up to 0.5 cm in diameter, is perthitic microcline that contains crosscutting veinlets or rims of granular oligoclase. The K-feldspar is strongly sheared and commonly shows a mortar structure. The type Tyringham (Stops 1 and 2) contains minor interlayers of mafic material. However, over 90 percent of the exposures are granitic gneisses.

Washington Gneiss

Washington Gneiss (Emerson, 1898, p. 20; 1899, p. 34) is named for distinct blue-quartz-bearing graphitic gneiss and biotite gneiss exposed near Washington, Massachusetts (East Lee quadrangle). The formation contains a mixture of rock types that grade into one another. The most abundant lithic type is a coarsely-ribbed, rusty weathering, blue-quartz biotite gneiss and schist (Stop 5). Layers 0.5 to 1 cm thick of bluish quartz "pebbles" are compressed in the plane of the foliation. The blue color is evidently a result of minute rutile crystals that can be detected in thin section under high magnification. Locally a white, mica-poor, plagioclase-rich blue-quartz-bearing granulite is interlayered with the more typical gneiss.

The Washington Gneiss passes laterally by interbedding of thin 2 to 3 foot mafic biotite-hornblende layers into massive amphibolites of the overlying hornblende gneiss unit.

Hornblende Gneiss

Hornblende-bearing gneisses and massive hornblende-garnet-sphene amphibolites commonly with small amounts of plagioclase (5 to 15 percent) grade into the Washington Gneiss, with which they are partially equivalent in age. The major part of the unit is either massive black-and-white-spotted hornblende-plagioclase granulite or well layered dark-gray hornblende-quartz-plagioclase-biotite gneiss. This map unit may correspond in position to that of the Lee Gneiss of Emerson (1898, p. 20; 1899, p. 33). The Lee Gneiss is also a mafic unit but differs from the amphibolites mapped on Beartown Mountain in that it contains abundant quartz grains that weather out on the surface giving the rock a distinctive quartz chaining. Both of these units (Lee and hornblende gneisses) may represent metamorphosed mafic volcanic rocks, perhaps flows interlayered with more felsic volcanic material.

Biotite Gneiss

Well layered, biotitic, quartz-rich paragneisses, feldspathic quartzites, and fine-grained magnetite-spotted granitic gneisses are exposed along the periphery of Beartown Mountain and over broad areas to the south. Locally the unit contains interlayers of calcite-chondrodite marble, diopside-hornblende calc-silicates (Stop 6), coarsely crystalline calcite-green diopside marbles, and thin biotite-rich mafic layers. The correlation of these biotite-rich gneisses, originally assigned to the Becket Gneiss (Emerson, 1899), is uncertain owing to Emerson's redefinition of the Becket (1917, p. 154) as a meta-intrusive of Precambrian age. Clearly much of what has been mapped as Becket Gneiss (Emerson, 1917) is metasedimentary; therefore the term is not used.

The abundant variation and repetition of lithic types within these biotitic gneisses makes this part of the Precambrian section the most difficult stratigraphic problem to solve. This problem is complicated by lithic similarity between some of the schistose rocks and the two-mica schists of the overlying Dalton Formation.

Cover Rocks

Dalton Formation

The Dalton Formation of probable Early Cambrian or Late Precambrian age consists of a heterogeneous sequence of tan weathering, feldspathic metaquartzites, two-mica quartz schists characterized by an abundance of black tourmaline, and less commonly schistose and quartz cemented meta-quartz pebble conglomerates. Strong lateral and vertical facies changes characterize this formation. With interbedding of clean, vitreous quartzite beds, it passes into the Cheshire Quartzite, with which it is laterally equivalent.

Within the area of Figure 2 the Dalton overlies with angular discordance the following Precambrian map units: Tyringham Gneiss, Washington Gneiss, and biotite gneiss. Where the contact can be closely located, the basal part of the Dalton is a very micaceous two-mica schist with irregular white milky quartz knots (Stop 2), or a greenish-gray two-mica gneiss with abundant detrital allanite grains. The latter member is best exposed at Umpachene Falls (Stop 10).

Cheshire Quartzite

Massive, white- to pinkish-tan weathering, vitreous meta-quartzite characterizes the Cheshire Quartzite of Early Cambrian age. Typical examples contain greater than 96 percent quartz and less than 1 percent total feldspar. Muscovite, tourmaline, magnetite, and zircon account for the remaining percentages.

The name Cheshire is used here only for those exposures with greater than 50 percent vitreous quartzite. The Cheshire is 500 to 800 feet thick in the vicinity of Great Barrington and Stockbridge, but it thins to a feather edge in an easterly direction. The exposures at Stop 10 (Umpachene Falls) illustrate these stratigraphic relations. A rusty weathering actinolite-rich zone marks the transition from Cheshire up into the overlying Stockbridge Formation (Stops 7-10).

Stockbridge Formation

Seven lithic subdivisions of the Stockbridge Formation proposed by Zen (1966) have been mapped in the area of Figure 2. The map units are given letter designations a through g and are grouped into three members on the generalized map (Fig. 2).

For a discussion of the stratigraphy of the Stockbridge Formation see Zen, Trip 3. The lithic subdivisions are briefly summarized below.

OCsg, medium- to dark-gray calcite marble with interbeds of cream to beige weathering dolostone.

OCsf, tan to gray weathering, crossbedded, sandy-textured dolostone or calcite dolostone.

OCse, coarsely crystalline, white to light-gray, blue-gray and white mottled, and massive white calcite marble that commonly is quarried.

OCsd, beige weathering sandy dolostone, or punky weathering calcitic metasandstone locally crossbedded and with white vitreous quartzite interbeds 1 to 3 cm thick. At metamorphic grades higher than the staurolite isograd, white diopside phlogopitic calc silicates are common.

OCsc, massive, light-gray weathering, steel-gray, very fine-grained calcitic dolostone with milky-white quartz knots 1 to 2 cm thick near top of unit.

OCsb, beige to light-cream weathering non-calcitic dolostone with abundant interbeds of punky weathering quartzites and silvery-gray phyllitic partings throughout much of the unit.

OCsa, massive to bedded, white to light-gray, crystalline dolomitic marble generally free of phyllitic or siliceous impurities. Base is gradational with underlying Cheshire Quartzite.

Stockbridge units a and b (Stop 3) and c, d, and e (Stop 9) will be seen on this trip. At grades higher than staurolite grade, east of East Mountain (Fig. 2), tablets of white diopside up to 1 cm long mark unit d and the upper part of unit c.

Walloomsac Formation

The Walloomsac Formation of Middle Ordovician age (Zen, 1966) is either a black, biotite-rich quartzose schist, or an orange weathering, schistose marble within the area of Figure 2. The schistose marble forms the base of the unit and regionally thickens to 200 to 300 feet in the western parts of the Stockbridge and eastern parts of the Egremont and Bashbish Falls quadrangles. The base of the formation marks a regionally developed unconformity (see Stop 4, Trip 2). Extensive exposures of this unit appear to coincide with areas of maximum denudation on the pre-Walloomsac erosion surface, suggesting that the carbonate-rich facies of the Walloomsac was in part derived from erosion of the Stockbridge Formation.

This basal unit will be seen at Stop 3 on the northeast end of Cobble Hill, where it is found beneath a thrust of the Precambrian gneiss.

The stratigraphy of the overlying Everett Formation, part of the Taconic allochthon, is discussed in Trip 2. The unit does not crop out within the area covered by Figure 2.

MAJOR UNSOLVED PROBLEMS

At the time of this writing several important structural problems are unsolved.

1. The location of the root zone of the Beartown Mountain nappe is not known. Attitudes of minor folds thought to be related to the thrusting appear to indicate movement from the northeast to the southwest. North of the Tyringham Valley (Stops 1 and 2), however, the Precambrian-Dalton contact is right-side-up, suggesting that these rocks are not a part of the Beartown Mountain slice because the Precambrian-Dalton contact is inverted on Beartown Mountain. Perhaps the root zone lies covered by rocks moved southwest on the fault north of Tyringham known as the Tyringham fault (Emerson, 1899).

2. It is now known that the Beartown Mountain slice is only one of a series of low angle thrusts that involve Precambrian rocks. The Precambrian front at this latitude appears to be marked by several overlapping thrust slices. These overlapping thrusts, such as the one at Halls Hill in the Monterey quadrangle, further complicate the geology by covering critical exposures of the underlying Beartown Mountain slice. Mapping in progress in the South Sandisfield quadrangle by David Harwood and in the Monterey and East Lee quadrangles by the writer should shed some light on this problem.

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ROAD LOG

Field trip stops and route will be in the East Lee, Stockbridge, Monterey, Great Barrington, and Ashley Falls quadrangles.

From Albany take Exit 2 of Massachusetts Turnpike at Lee, Massachusetts. Head south on Rte. 102 toward South Lee-Stockbridge. Immediately turn left (0.1 mile) onto Tyringham Road. Log starts from this point.

Mile

0.0 Intersection Rte. 102-Tyringham Road, Lee, Mass., East Lee quadrangle. Head southeast toward Tyringham.

0.9 Discontinuous outcrops of Dalton Formation and Cheshire Quartzite left side of road. Right-side-up west-dipping sequence; valley to right exposes units a, b, and c of Stockbridge.

0.5 Intersection Meadow St., view to right of north slopes of Beartown Mt.

2.0 Stop 1. Exposure left side road of type Tyringham Gneiss (Emerson, 1899) with interlayered dark-gray, well-layered biotite gneiss. In field well-layered dark-gray biotite gneiss strikes N.35°E. and dips steeply northwest. Two later foliations, N.30°E.45°S.E. and N.25°W.30°N.E., cut this layering. At east edge of field large crops of massive granitic gneiss typical of the bulk of the Tyringham. Note the indistinct banding and well developed foliation. Rock is a microcline-plagioclase-quartz-muscovite-biotite gneiss. Compositional layering is intruded by K feldspar quartz pegmatites, and that is cut by the N.20°E.45°S.E.-dipping foliation. The foliations and pegmatites are thought to be of Precambrian age because the Dalton (Stop 2) unconformably overlies rocks showing these features.

Continue southeast on Tyringham Road.

3.4 Pass house of Hansel and Gretel on left; slow down for hidden drive.

3.6 Turn left on George Canon Road.

3.8 Stop 2. Park at cabin on left. Walk up hill to stream crossing. Tyringham-Dalton contact relations (unconformity) and lithology of basal Dalton. We will hike up the hill to first outcrop, then view the rocks on our descent, walking up section.

Stop 2-a. Outcrop of Tyringham Gneiss, compositional layering N.20°E.80°S.W., oblique intersecting foliations N.20°W.45°S.W. and N.40°E.15°S.E.

Stop 2-b. Small outcrop of Tyringham in stream. Compositional

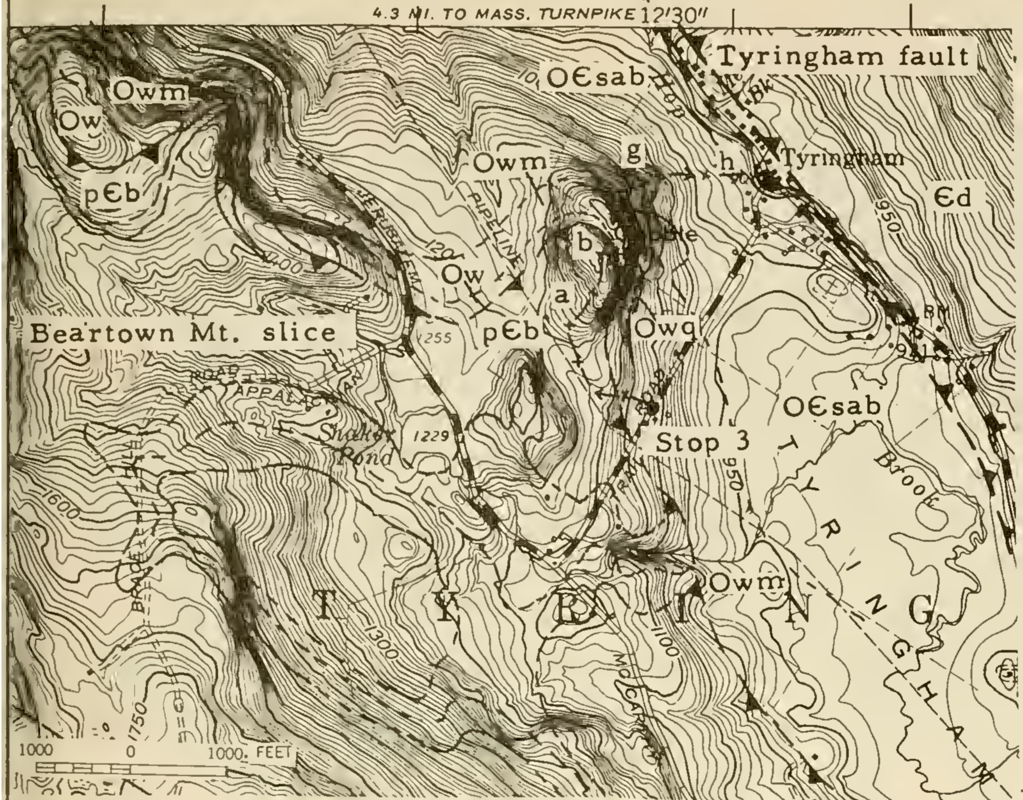
layering still trending northeast-southwest; note linear structure.
Stop 2-c. Small waterfalls by well house. Dark biotite-rich schist with muscovite and white quartz knots and black tourmaline. Bedding and parallel foliation, N.50°W.25°S.W., is folded by late folds that plunge S.15°W. The unconformable relations seen here are duplicated along the entire Dalton-Tyringham contact north of Tyringham. Conglomerates are locally developed, but characteristically not right at the unconformity. The basal unit is commonly a schistose rock, although locally thin (10 to 20 feet thick) beds of arkosic quartzites may mark the base of the Dalton. Black tourmaline is a never-failing accessory mineral.
Stop 2-d. In stream, flaggy, impure quartzites typical of the Dalton. Note well developed bedding and highly feldspathic layers. Continue downstream to road and cross into brook below road.
Stop 2-e. Interbeds of vitreous quartzite, tentatively assigned to the Cheshire. Return to bus.

4.0 Return to intersection of George Canon Road and Tyringham Road. Turn left. View to right of Cobble Hill (Stop 3).

4.7 Town of Tyringham; turn right on Jerusalem Road (no signpost).

4.8 Bear right at Y; head up hill. View of Cobble Hill at right (west). Stockbridge at base, overlain by Walloomsac, and finally gneiss caps the hill.

5.2 <u>Stop 3. Cobble Hill traverse - Precambrian biotite gneisses of Beartown Mountain slice resting on Walloomsac.</u> We will walk to the top and look at the rocks on the way down the steep northeast slope, and will rejoin bus at the town of Tyringham (see Fig. 3).
a. Outcrops in field of rusty weathering, well-layered biotite gneiss with minor amphibolite layers. These gneisses are believed to belong to the youngest of the Precambrian units. Note the differences and similarities with the schistose rocks of the basal Dalton.
b. Walk to crest of hill; note late southeast-plunging folds that fold the early foliation.
c. At base of cliff on northeast edge of the Cobble. Overhanging cliff of gneiss exposing a large nearly recumbent fold in gneiss. The fold plunges S. 20° E. at 5° with a northeast over southwest rotation sense (clockwise). Note an earlier set of folds with a counterclockwise rotation sense and folded lineation. The large recumbent folds probably formed at the time of emplacement of the Beartown Mt. nappe.
d. Walk carefully along cliff to an infold of epidote-actinolite-quartz pebble conglomerate. This could be basal Dalton or Walloomsac infolded at the base of the gneiss; but age assignment of this lithology is uncertain, as the rock has not been found elsewhere.
e. Walk down cliffs to northeast to exposures of Ow (schist and schistose



Explanation

Ow

Walloomsac Fm. (schist)

Owm

Schistose marble of Walloomsac Fm.

Owq

Quartzite in Owm

OЄsab

Stockbridge Fm., units a and b

Єd

Dalton Fm.

pЄb

Biotite gneiss

Thrust fault, teeth on upper plate

Fig. 3. Geologic map showing Stop 3 (Cobble Hill traverse), Monterey quadrangle. Contacts within Beartown Mt. slice are members of the Biotite gneiss unit and are not described individually here. Letters a, b, g, and h denote stops along traverse.

Explanation

Ow

Walloomsac Fm.

OЄsf
OЄse
OЄsd — Stockbridge Fm.
OЄsc
OЄsb

1000 0 1000 FEET

Fig. 4. Geologic map showing the location of Stop 9, Great Barrington quadrangle. Unit d of the Stockbridge Fm. is shaded.

marble) that underlies the biotite gneiss. Pasture contains scattered outcrops of Owm and Ow.

f. At edge of hill quartzite bed in Owm. Beds of quartzite are relatively rare in the Walloomsac and are only found in areas of maximum erosion on the Middle Ordovician unconformity. Perhaps the quartz in these quartzites was locally derived by erosion of the Cheshire or Dalton.

g. Continue down the cliffs in Owm and finally into dolomitic marbles of the Stockbridge. The impure beige weathering dolostones of unit b of the Stockbridge are found at the base of the hill.

h. At Hop Brook, a small exposure of calcitic dolostone with tourmaline and golden yellow phlogopite. This exposure was called Precambrian by Emerson (1899) because of reported chondrodite in the marble. However the condrodite here has not been confirmed. The rock closely resembles either unit b or c of the Stockbridge Formation and does not resemble other Precambrian marbles (Stop 7) that do contain chondrodite. Marbles in the Precambrian here are largely calcitic, although dolomitic chondrodite marbles are known from several localities. The floats of chondrodite marble referred to by Emerson (1899) can still be seen at the base of the crops. Return to bus at Tyringham center.

5.7 Turn left on Tyringham Road. Re-enter East Lee quadrangle.

8.4 Turn left onto Meadow Street and cross broad Tyringham Valley underlain by Stockbridge units c, b, and a. Enter Stockbridge quadrangle.

9.3 Turn right on Fernside Road. Hill to left duplicates Cobble Hill relations: Stockbridge a, b at base; Walloomsac overlain by biotite gneisses at top of first cliff.

10.3 One of Berkshire's scenic dude ranches. Ugh!

10.6 Turn right after crossing railroad track into Pine Street.

10.9 Bear left at Y onto Beartown Mt. Road. Walloomsac exposures to the west.

11.2 Feldspathic quartzites of the Dalton Formation, dip to south toward Beartown Mt. and overlie the Walloomsac as part of the Beartown Mountain slice.

11.4 Precambrian biotite gneisses, compositional layering and foliation dip south, thrust over the Dalton.

12.1 Marvelously dirty pig sty on left. We are driving on the biotite gneiss and the hornblendic gneiss units, approaching the hingeline of Beartown Mt. nappe.

12.5 Stop 4. At BM 1363 (Stockbridge quadrangle). Roadcut and cliffs of Tyringham Gneiss in the core of the Beartown Mt. nappe. East-west nearly vertical compositional layering is intersected by two cleavages that dip gently north and south, producing rhombic cleavage fragments and a strong east-plunging lineation. Zircons from this exposure and from the next are being dated by Robert Zartman of the U. S. G. S. At the time of this writing, the data are not yet available. When completed, these will be the first ages from the gneisses of the Berkshire Highlands. To the east the map relations indicate this unit closes as a west-plunging synform, thus ruling out a simple anticlinal structure.

LUNCH STOP

Continue south on Beartown Mt. Road; enter Great Barrington quadrangle.

13.2 Intersection on left of road to Mt. Wilcox. We are crossing the hingeline of the nappe; the section repeats going south.

13.6 Low exposure to right of schistose rock of the Washington Gneiss, and associated hornblende amphibolites, massive hornblende garnet amphibolites are exposed south of this point on the hill to the west.

14.4 Small bridge; everybody out if necessary. This point marks the axis of the late southeast-plunging synform that folds the early anticlinal nappe.

15.8 Stop 5. Washington Gneiss (Great Barrington quadrangle). We have crossed the hornblendic gneiss and the Tyringham gneiss, and now are on the southwest flank of Beartown Mt. in a section that is thought to top toward the southwest. Exposures on left (east) side of dirt road are rusty weathering, blue-quartz, plagioclase, biotite gneiss. Interlayers of thin biotite-rich amphibolite layers are isoclinally folded. Walk west into woods to weathered exposures of the same rock showing the distinctive blue-quartz ribbing and the white granulite lithology. Minerals in the Washington include muscovite, biotite, hornblende, plagioclase, quartz, and scapolite. Locally graphite-rich schists and actinolite-bearing calc silicates are found. Note muscovite-biotite schist at roadcut. Reboard bus. Continue south on Beartown Mt. Road.

16.2 At bend in road to southeast and steep decline, contact biotite gneiss unit.

17.3 Benedict Pond on left. Fine-grained granitic gneiss of the biotite gneiss unit. Turn right at Y onto new road that follows A.T. on topographic sheet.

17.7 Turn left on Stony Brook Road.

17.8 Stop 6. Calc silicates in biotite gneiss unit on southwest flank of Beartown Mt. Please DO NOT HAMMER at rocks in outcrop. There is enough float of everything to go around. Walk east from field into woods. Cliffs of well layered hornblende-plagioclase and biotite plagioclase quartz gneiss with minor layers of calcite-chondrodite marble, diopside-hornblende calcite marble and minor layers of plagioclase granulite with large blotches of green hornblende and 1/4-inch chocolate-brown sphene crystals. In the vicinity of crosscutting pegmatites, chlorite pseudomorphs after hornblende or diopside can be seen. This distinctive unit is thin but does not appear to be stratigraphically persistent, although rocks of this kind are good marker horizons in gneisses at Benton Hill (Fig. 1) and in the South Sandisfield quadrangle, according to David Harwood. Note dips to the northeast. The distant view to the southwest looks over the Lake Buel window (underlain by Stockbridge) to East Mountain, where biotite gneisses with calc silicates like this and the hornblendic gneiss unit rest with thrust contact on the Dalton Formation marking the southwest edge of the Beartown Mt. slice.

Continue southeast on dirt road.

19.4 Turn right on Brett Road.

19.9 Turn right on Rte. 23. Exposures of recumbently folded Stockbridge units c, d, e. Rocks that are beneath the Beartown Mt. slice. Note large tablets of white diopside.

21.0 Intersection of Rte. 57. Continue west on Rtes. 23 and 57.

21.1 Stop 7. Overturned Stockbridge and Cheshire at base of Beartown Mt. slice. Walk up from road to yard by small house. Massive Cheshire overlies Stockbridge unit a, with thin, rusty weathered actinolite zone marking the contact. Note northeast over southwest rotation sense of folds.

Continue west on Rte. 23 and 57 past low road cuts of flat lying Stockbridge in the Lake Buel window.

22.2 Turn left on Lake Buel Road.
Stop 8. Dalton Formation of Monument Mt. slice at northeast face of East Mt. Walk into woods to large cliffs of isoclinally recumbently folded feldspathic quartzites of the Dalton Formation. Rotation sense northeast over southwest. Precambrian gneiss like those at Stop 6 overlies these rocks farther up the hill.

Continue southeast on Lake Buel Road.

24.9 Turn right at Y onto Mill River Road.

26.8 Take right. Turn at Y onto Sheffield Road.

27.0 Turn right into yard of yellow farmhouse by burned barn.
Stop 9. Exposures of extremely folded and refolded carbonates, units c, d, e of the Stockbridge. Lake Buel window. You will walk from the house, starting in white calcite marbles of unit e, to rusty weathering diopside, calc silicate rock with interlayers of white quartzite exposed behind the barn (unit d), and finally fissil dolostones of unit c in pasture past the fence. Note the complex fold patterns. This type of deformation is characteristic of the rocks in the Lake Buel window. Please don't expect your leader to explain the folds; have fun for yourself. The overall map pattern at this locality can be seen in Fig. 4.

Return to bus, head east (left turn) onto Sheffield Road.

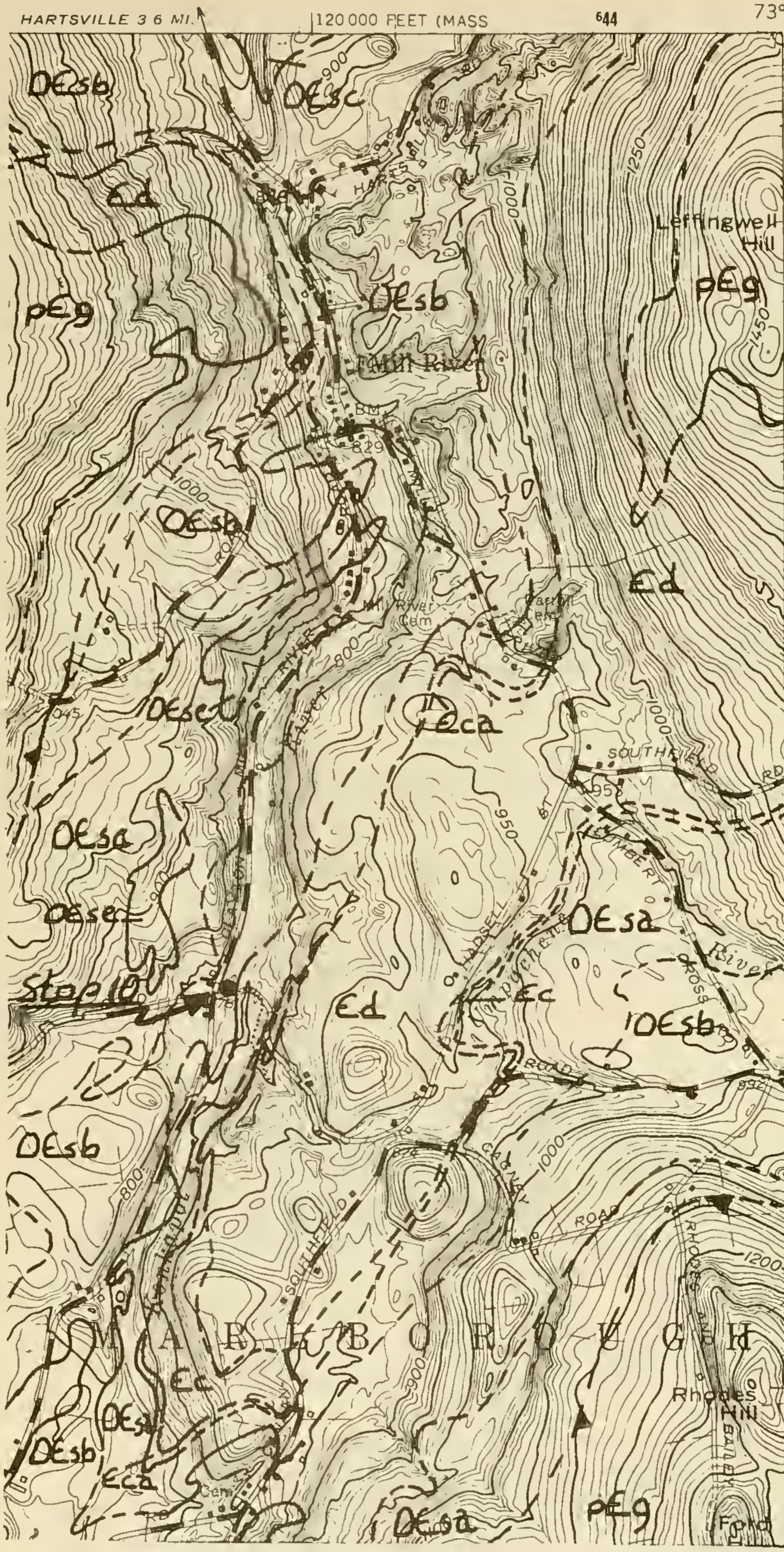
27.4 Turn right onto Mill River Road.

28.0 Enter Ashley Falls quadrangle. (See Fig. 5 for location of route to Stop 10.)

28.9 Cross Konkapot River and bear right. Exposures Stockbridge units a and b in river.

29.2 Mill River; turn right on Clayton Road; cross river and turn left down Clayton Road.

30.5 Turn left on dirt road to Umpachene Falls.
Stop 10. Ashley Falls quadrangle. Stockbridge unit a, Cheshire Quartzite, and Dalton Formation of the autochthon. Exposures of white crystalline dolomitic marble, Stockbridge unit a, with gentle east-dipping foliation will be seen on the way to the falls. At base of falls a rusty weathering actinolite zone marks the Cheshire-Stockbridge contact, overlain by a quartzite bed approximately 10 feet thick (Cheshire). Up stream this quartzite passes into silvery gray gneissic rock containing irregular white clots of feldspar and quartz. This rock was previously mapped as Precambrian Becket Gneiss (Emerson, 1917). The Cheshire returns upstream and now overlies the gneissic rock in a small syncline. The major structure is an anticline with nearly horizontal axial plane foliation that is locally warped in gentle folds. This deformation style is characteristic of the overlying parautochthonous Precambrian rocks at Benton Hill to the east (Fig. 1). This exposure illustrates several important stratigraphic relations:
a) The Cheshire Quartzite that measures 500 to 800 feet in thickness in the Great Barrington quadrangle, is here only 10 feet thick, suggesting very rapid thinning of the quartzite facies eastward. Locally in the



Explanation

Ow
Walloomsac Formation

OЄse
OЄsd
OЄsc
OЄsb
OЄsa
Stockbridge Formation

Єca
Rusty weathering actinolite rock

Єc
Cheshire Quartzite

Єd
Dalton Formation

pЄg
Precambrian gneisses, here not differentiated

1 ½ 0
MILE

Fig. 5. Geologic map showing the location of Stop 10, Umpachene Falls (northeast corner of the Ashley Falls quadrangle).

area around Umpachene Falls, the Cheshire is absent altogether and schistose rocks of the Dalton are in contact with unit a of the Stockbridge.
b) The Umpachene type of Dalton seen here differs markedly from the typical orange-tan weathering feldspathic quartzites of the Dalton (Stops 2 and 8).
Evidently the Dalton Formation is marked by rapid vertical as well as lateral facies changes. The relations seen here indicate the feldspathic and gneissic facies of the Dalton are lateral equivalents of the clean quartzite facies (Cheshire). In areas to the east Cheshire-equivalent rocks may be gneissic or schistose rocks previously assigned to the Becket or other Precambrian map units.

Trip 2

STRATIGRAPHY AND DEFORMATIONAL HISTORY OF ROCKS OF THE TACONIC RANGE NEAR GREAT BARRINGTON, MASSACHUSETTS*

by

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INTRODUCTION

This field trip will deal specifically with the stratigraphy of rocks of the Chatham and Everett slices of the Taconic allochthon and their relation to the underlying autochthonous rocks. An attempt will be made to decipher the complex effects of four recognized Paleozoic deformational episodes and their relationship to periods of metamorphism. The trip will follow a west to east route, progressing from rocks of chlorite grade in the western part of the State Line quadrangle to those of staurolite grade near Stockbridge, Massachusetts. All stops will be in the State Line and Stockbridge quadrangles.

The State Line and Stockbridge quadrangles (Fig. 1) are located on the eastern edge of the Taconic Range where one can determine relations between the allochthonous rocks of the Taconic sequence and the underlying autochthon, which ranges from probable Early Cambrian to Middle Ordovician in age. Autochthonous rocks include, from the base, the Dalton Formation, Cheshire Quartzite, the Stockbridge Formation, and the Walloomsac Formation. The carbonate rocks of the Stockbridge are overlain by schistose rocks previously termed the Berkshire Schist (Dale, 1923). In recent mapping in southwestern Massachusetts and adjacent New York and Connecticut (Zen and Hartshorn, 1966; Zen and Ratcliffe, in press; Ratcliffe, 1965; Ratcliffe, unpub. data; Ratcliffe and Burger, unpub. data) this unit has been subdivided into two major units: (1) the Walloomsac Formation, a lower dark colored schist that contains Middle Ordovician fossils, and (2) an upper heterogeneous sequence of green, gray-green, or purple phyllite believed to be allochthonous rocks of the Taconic sequence. The results of these investigations are summarized in Figure 1. The bulk of the allochthonous rocks are assigned to the Everett slice (Zen, 1967), but the northwest corner of the State Line quadrangle is underlain by rocks of the Chatham slice of Zen (1967).

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*Publication authorized by the Director, U. S. Geological Survey

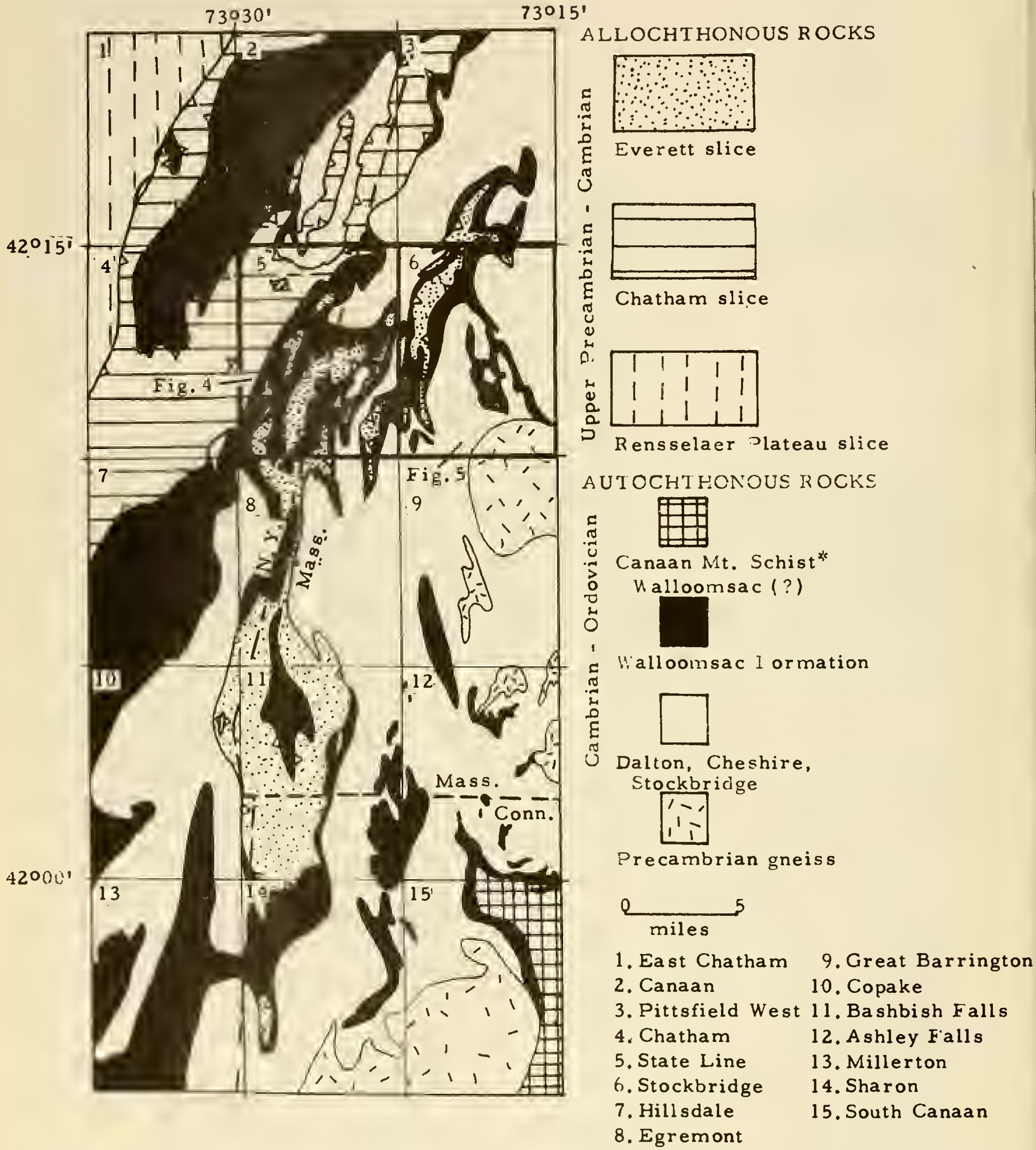


Fig. 1. Generalized geologic map of southwestern Massachusetts and adjacent areas of Connecticut and New York. Diagrams show the distribution of the allochthonous Taconic rocks. For more detailed map see Trip 1, Fig. 1. Data from Zen (1967), Zen and Hartshorn (1966), Zen and Ratcliffe (in press), Ratcliffe (1965), Ratcliffe and Burger (unpub. data), and Fisher et al. (1961). *Canaan Mountain Schist of Rodgers and others (1959).

The Chatham slice in the State Line quadrangle discordantly overlies map units of the Stockbridge Formation. Moreover, map units within the Chatham slice are discordant with the base of the slice.

The Everett slice, exposed to the east, overlaps the older Chatham slice. The two slices are not in contact because they are separated by a belt of parautochthonous Walloomsac. This intensely deformed belt of parautochthonous Walloomsac contains blocks of Stockbridge carbonates in a tectonic breccia formed at the base of the higher Everett slice (Zen and Ratcliffe, 1966).

ACKNOWLEDGEMENTS

The work presented here was begun in 1962 under the guidance of Dr. R. H. Jahns, then of the Pennsylvania State University, and with the financial support of the Penrose Fund. Subsequent funding was provided from 1966 to the present by the U. S. Geological Survey in cooperation with the Massachusetts Department of Public Works. My interest in Taconic geology was initially whetted by John MacFadyen, Jr., of Williams College while I was an undergraduate there. During the summer of 1961 I had the opportunity of working with E-an Zen on the Bashbish Falls quadrangle. I am very grateful to him for help in all phases of this study, for his willingness to share ideas, and for his helpful criticism.

STRATIGRAPHY

The stratigraphic column is divided into two parts: a sequence of eugeosynclinal rocks believed to be entirely allochthonous (Nassau and Everett Formations of Late Precambrian or Early Cambrian age), and a quartzite, carbonate, and pelite sequence largely autochthonous but in part parautochthonous. The latter consists of the Dalton Formation, the Cheshire Quartzite, and the Stockbridge and Walloomsac Formations and is believed to range in age from Early Cambrian to Middle Ordovician. Approximate thickness and correlation of each of these rocks with rocks of other areas is indicated in Figure 2. The explanation and geologic map of the State Line and Stockbridge quadrangles are presented in Figures 3, 4, and 5. Location of trip stops is shown in Figures 4 and 5.

Autochthonous Rocks

Dalton Formation - Cheshire Quartzite

White to pinkish-gray weathering, massively bedded vitreous quartzite is characteristic of the Cheshire Quartzite. With increase of feldspathic quartzites and muscovite-biotite schists, the Cheshire passes gradationally into the underlying Dalton Formation. For a discussion of the Dalton Formation see Trip 1.

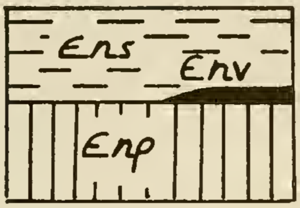
Age	Pine Plains area, N. Y. (Knopf, 1962)	Bashbish Falls quadrangle Conn.-Mass.-N. Y. (Zen and Hartshorn, 1966)	State Line quadrangle and vicinity, this report	
E. Cambrian or Uncertain		Everett Formation	Everett Formation	1000'
			Nassau Formation	2500'
Middle Ordovician	Normanskill Shale	Walloomsac Formation= Egremont Phyllite	Walloomsac Formation	2000'
			Schistose marble Owm	0-100'
	Balmville Ls.	Limestone interbedded with schist	Limestone Owl	0-50'
Early Ordovician	Copake Limestone	Stockbridge Formation: Unit g	Stockbridge Formation: Unit g	400'?
	Rochdale Limestone	Unit f	Unit f	0-150'
	Halcyon Lake Fm.	Unit e	Unit e	100-300'
Middle and Late Cambrian	Briarcliff Dolostone	Unit d	Unit d	0-100'
		Unit c	Unit c	800'
Early Cambrian	Pine Plains Fm.	Unit b	Unit b	600'
	Stissing Dolostone	Unit a	Unit a	600'
	Poughquag Quartzite	Cheshire Quartzite	Cheshire Quartzite	0-700'
Cambrian ?		Dalton Formation	Dalton Formation	700-1000'
Pre-cambrian			Tyringham, Becket, Washington Gneisses	?

Figure 2. CHART SHOWING CORRELATION OF MAP UNITS IN THE STATE LINE, STOCKBRIDGE, AND GREAT BARRINGTON QUADRANGLES WITH ADJACENT AREAS.

Late Precambrian to Early Cambrian

ALLOCHTHONOUS ROCKS

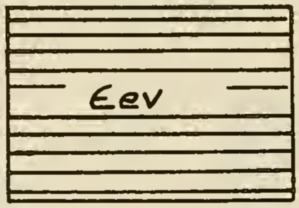
Chatham slice (west)



Nassau Formation

Ꞓns - pale-green, gray metasiltstone and phyllite including massive graywacke (Rensselaer(?))
Ꞓnv - metavolcanic rocks
Ꞓnp - purple and green phyllites with metaquartzite beds

Everett slice (east)



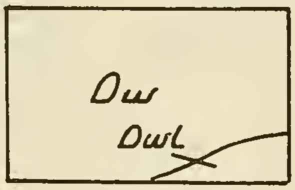
Everett Formation

Ꞓev - green, gray, lustrous phyllite, and siliceous phyllite, containing minor beds of graywacke and blue-quartz pebble conglomerate

AUTOCHTHONOUS ROCKS

Mid Ordovician

Ordovician



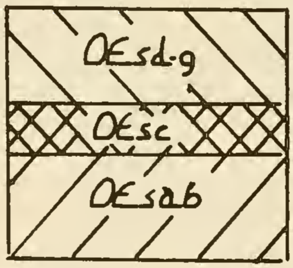
Walloomsac Formation

Ow - black, graphitic phyllite with fossiliferous limestone (Owl), or schistose feldspathic marble (Owm) near base

Unconformity

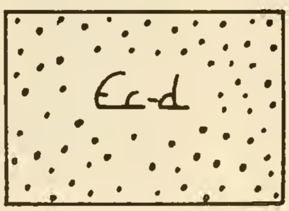
Early Cambrian to Early Ordovician

Cambrian



Stockbridge Formation

OꞒsd to g - upper calcitic units; OꞒsc - calcitic dolostone unit C; OꞒsab - combined units A and B, lower dolomitic units



Cheshire Quartzite - Dalton Formation

Ꞓcd - feldspathic, schistose and pure vitreous quartzites.

Unconformity

Precambrian gneisses

Fig. 3. Explanation for geologic maps, Figures 4 and 5.

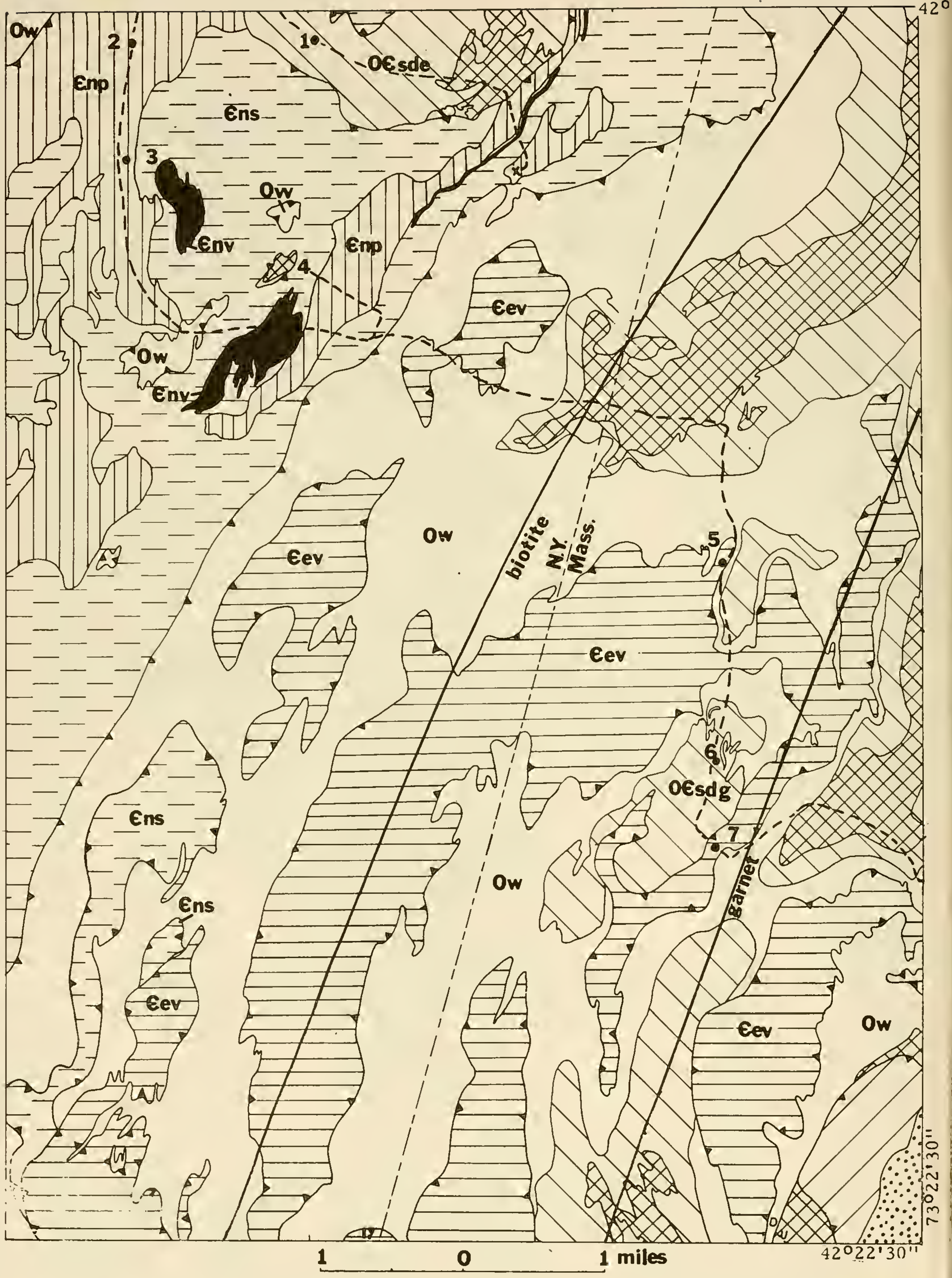


Fig. 4. Geologic map, State Line quadrangle, showing location of stops 1-7.

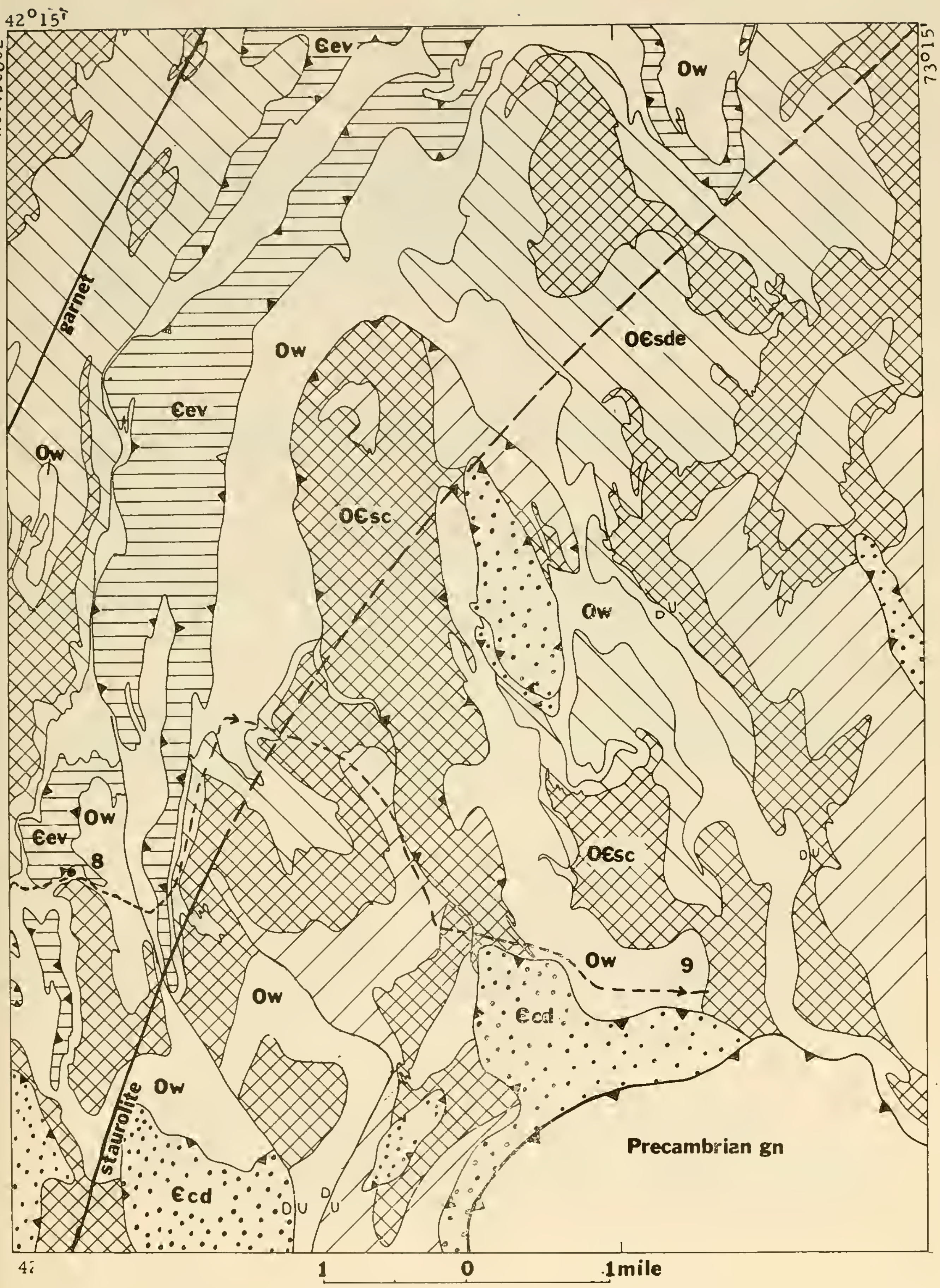


Fig. 5. Geologic map, Stockbridge quadrangle, showing location of stops 8, 9.

In mapping, only exposures characterized by massive vitreous quartzites were assigned to the Cheshire Quartzite.

The upper contact with the Stockbridge Formation is also gradational, illustrating depositional continuity from the impure feldspathic metasedimentary rocks of the Dalton Formation up into the carbonate rocks of the Stockbridge Formation. The Cheshire is assigned an Early Cambrian age on the basis of Olenellus fragments found in feldspathic metaquartzites underlying the vitreous facies on the Dome near Williamstown, Massachusetts (Walcott, 1888, p. 235-236).

Stockbridge Formation

Zen (1966) has recently subdivided the Stockbridge into seven lithic units. These subdivisions have now been mapped in most of southwestern Massachusetts (see Fig. 3 and 4; and Zen, Trip 3, this volume). Minor facies changes are present even within this area. However, the major lithic units are remarkably persistent. On the basis of the overall lithic sequence and the Middle Ordovician age of the overlying Walloomsac Formation, the Stockbridge Formation ranges from Early Cambrian to Early Ordovician in age. The major lithic types are briefly summarized in Trip 1. Correlation with the carbonate section at Pine Plains, New York (Knopf, 1962) is given in Figure 2.

In Figures 4 and 5 the carbonate stratigraphy is grouped into an upper sequence (units g, f, e, d), unit c, and a lower sequence that is units b and a combined. These designations correspond with the usage on the maps in Zen, Trips 3 and 4 this volume.

Walloomsac Formation

Dark-gray to dull jet-black, fissile phyllite is characteristic of the Walloomsac Formation (Ow) (Zen, 1966). Lenses of limy schists, schistose marbles, and calcite marbles are sometimes found near the base of the formation (Owl) in exposures up to 20 feet thick. This limestone carries abundant fossils in exposures at No Bottom Pond (Stop 4) and several other localities in the western part of the State Line quadrangle. The Walloomsac Formation is assigned a Middle Ordovician age (Zen, 1966). The basal carbonate-rich zone of the Walloomsac appears to thicken to the east, where it is an impure feldspathic calcite marble or schistose marble at least 250 feet thick. The thickening of this calcitic member eastward is regionally characteristic in southwestern Massachusetts. The increased feldspar content probably is derived from the feldspathic Precambrian gneisses and Dalton Formation exposed to erosion during Middle Ordovician time.

The base of the Walloomsac marks a major break in the stratigraphic section, a widespread unconformity that bevels all units of the Stockbridge Formation. Physical evidence for an angular discordance can be seen at Wildcat Hollow in the Sharon quadrangle (Zen and Hartshorn, 1966; Zen, Trip 3, this volume) and at No Bottom Pond in the State Line quadrangle (Stop 4). The unconformable relations seen at this locality indicate the Stockbridge (unit c) was steeply dipping prior to deposition of the Walloomsac Formation.

Allochthonous Rocks

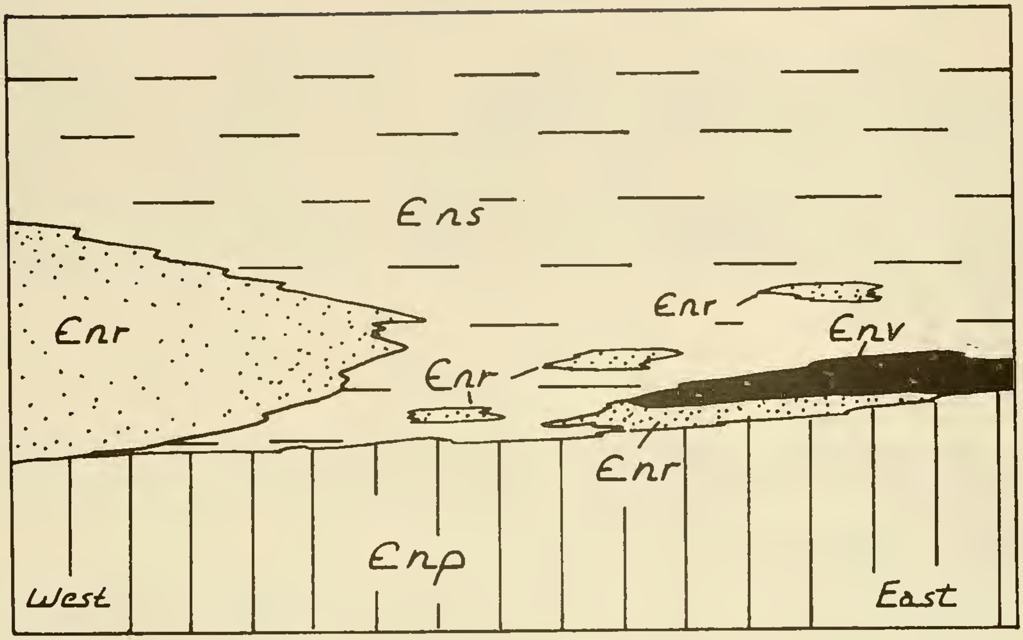
Rocks believed to be allochthonous are assigned to two separate thrusts, the Chatham slice in the west and the Everett slice in the east.

Nassau Formation (Chatham slice)

Allochthonous rocks found in the lowermost slice (Chatham slice) are assigned to the Nassau Formation of late Precambrian(?) to Early Cambrian age as used by Bird (1962) and Bird and Rasetti (1968). The Nassau crops out in Rensselaer and Columbia Counties, New York. In the State Line area the Nassau is composed of three main members: (1) a purple and green lower phyllite member (Ꞓnp), (2) a graywacke member (Rensselaer(?)) (Ꞓnr), and (3) an upper member, predominantly green-colored metasiltstone and phyllite (Ꞓns). Metavolcanic rocks (Ꞓnv), probably both flows and tuffs, occur near the middle of the section and are closely related to exposures of graywacke (Ꞓnr) and the purple phyllite (Ꞓnp)-green phyllite (Ꞓns) contact.

The massive graywacke member is best exposed in the western part of the State Line quadrangle (Fig. 3) where the northern end of the Austerlitz outlier of the Rensselaer Graywacke (Balk, 1953) is exposed. To the east the graywacke occurs as lenses within the green metasiltstone-phyllite unit (Ꞓns). These relations indicate a thinning of the graywacke beds eastward, thus confirming the conclusion of Balk (1953) and Bird (1962) that the graywacke (Rensselaer(?)) had a westerly source. The facies relations within the Chatham slice in the State Line area are pictured in Figure 6.

The metavolcanic rocks (Ꞓnv) are interpreted as flows and volcaniclastic deposits rather than as hypabyssal intrusions because of their persistence near the Ꞓnp-Ꞓns contact, the concordant or gradational contacts with Ꞓns, and the amygdaloidal borders. Balk's report (1953) of clasts of metavolcanic rock identical to Ꞓnv within the Rensselaer Graywacke of the Austerlitz outlier supports this interpretation.



Ɛns: Green-gray to gray metasiltstone or phyllite with 1 to 3 cm beds of greenish meta-quartzite and olive drab, fine-grained metasiltstone; pale-green phyllite with minor quartz conglomerate lenses and more rarely gneiss boulder conglomerate beds, and dark-green graywacke beds as much as 10 cm thick.

Ɛnr: Rensselaer Graywacke member, massive, bedded, dark- to pale-green graywacke or subgraywacke with minor blue quartz pebble, coarse gneiss boulder, and gneiss pebble conglomerate layers. Interfingers with and grades into massive green-gray to gray metasiltstones of Ɛns containing numerous lenses of graywacke too small to be shown on map.

Ɛnv: Metavolcanic rocks, dark-green to yellow-green, ilmenite-leucoxene-chlorite-actinolite-hornblende-epidote-plagioclase greenstone forming conformable, massively layered units as much as 10 m thick. Individual layers commonly show relict intersertal igneous texture grading toward finer grained, strongly foliated rock with scattered ankeritic amygdaloidal(?) fillings at contact with Ɛns or Ɛnr.

Less commonly the rock is dark-brown-green weathering stilpnomelane-plagioclase-amphibole-ilmenite-leucoxene meta-andesite(?) porphyry containing relict euhedral plagioclase phenocrysts up to 2 cm long; unit is fine grained and amygdaloidal near some contacts with Ɛns.

Ɛnp: Dark maroon phyllites, green and purple laminated or mottled phyllites, including red or green metashale chip conglomerate layers, and numerous light-green or purple-tinted meta-quartzites 1 to 10 cm thick that are commonly crossbedded. Light-green to gray weathering quartzite or subgraywacke forms lenticular bodies as much as 20 m thick at and near the top of Ɛnp.

Fig. 6. Schematic representation of facies relations within the Nassau Formation in the State Line quadrangle. (Not to scale.)

Everett Formation (Everett slice)

The Everett is primarily a greenish-gray phyllite or siliceous phyllite (Ꞓev). This phyllite contains abundant sericitic muscovite and an optically negative iron-rich chlorite whose index varies between 1.64 and 1.63. With increase in metamorphic grade, chloritoid is widespread and paragonite is sparingly developed. Magnetite commonly forms pea-sized metacrysts in the biotite metamorphic zone. Lithically the Everett closely resembles the upper unit of the Nassau (Ꞓns), but it lacks volcanic rocks and abundant interbedded purple phyllites. Massive graywacke similar to the Rensselaer Graywacke is recognized in the Everett, as are minor stilpnomelane-blue quartz pebble conglomerate lenses.

STRUCTURE

Taconic Thrusts

Rocks believed to be allochthonous and older than the bulk of the autochthonous sequence (Dalton, Cheshire, Stockbridge, Walloomsac) are found in two major overlapping thrust slices that occur in diagonal belts parallel to the regional northeast foliation-bedding trend.

The lowermost slice is coextensive with the Chatham slice of Zen (1967, plate 1) and is underlain in the northwest corner of the State Line quadrangle by either Walloomsac or Stockbridge. The thrust is breached in the carbonate valley near the New York State Thruway (Stop 1), exposing contact relations that indicate that the rocks of the Nassau Formation discordantly overlie units OꞒsc, OꞒsd, and OꞒse of the Stockbridge. Map units within the slice, moreover, are discordant to the contact, indicating a clearcut thrust relation. Autochthonous carbonates are exposed in a small window at No Bottom Pond (Stop 4). The eastern or trailing edge of the slice is covered by an overlapping belt of parautochthonous Walloomsac that probably formed beneath the Everett slice exposed to the east. Erosional outliers of parautochthonous Walloomsac rest with discordance on the Chatham slice one mile southwest of Stop 4. At this locality blocks of carbonate in a polymictic tectonic breccia float in a matrix of Walloomsac. Tectonic breccias of this kind are preserved at the sole of the Everett slice and at many places in the State Line and Egremont quadrangles (Zen and Ratcliffe, 1966).

The Everett slice rests directly on parautochthonous Walloomsac along its western edge and on autochthonous rocks along its eastern edge. Over broad areas in the Bashbish Falls quadrangle (Zen and Hartshorn, 1966), the Egremont quadrangle (Zen and Ratcliffe, in press), and the State Line and Stockbridge quadrangles the Everett rests discordantly on the Stockbridge Formation.

Tectonic Disturbances

Multiple non-coaxial folding is evident in all rocks in the State Line and Stockbridge quadrangles. The effects of four deformational episodes of Paleozoic age are recognized, the last two of which produced easily recognized foliations. The structural terminology used is listed below.

Tectonic event	Foliation	Tectonic style
D_0 (pre-Walloomsac)	None recognized	Folding, high angle faulting, (angular unconformity preserved).
D_1 (post-Walloomsac)	None recognized	Isoclinal recumbent minor folds perhaps intrafolial and related to Taconic thrusting of the Chatham and Everett slices.
D_2	Major foliation (S_2) formed, which trends NE, dips steeply E	Isoclinal flexural flow folds in all units. Thrust contacts folded.
D_3	(S_3) crenulation or slip cleavage formed, trending N. 15°W. and commonly vertical or west dipping. Locally N. 50°E. in the Chatham slice.	Chevron type, to strain slip folds common in schistose rocks. Intensity of folding increases to east. Locally S_3 develops a true foliation in exposures to the east.

Pre-Walloomsac Deformation (D_0)

The base of the Walloomsac Formation marks a widespread unconformity that bevels all units of the Stockbridge. Erosion appears to have been greatest in areas closest to the present Precambrian front. At a locality in the Ashley Falls quadrangle the Walloomsac rests on O€sa of the Stockbridge 25 feet from the Cheshire-Stockbridge contact. Physical evidence for angular unconformity can be seen at Wildcat Hollow in the northwestern corner of the Sharon quadrangle where an angular discordance of 10° was documented (Zen and Hartshorn, 1966). At the No Bottom Pond Window (Stop 4) fossiliferous limestone interbedded with black phyllites of the Walloomsac overlies unit c of the Stockbridge with approximately a 70° angle. At several localities in the State Line and Egremont quadrangles Walloomsac rests on inverted Stockbridge units as well as across a fault trace, suggesting that the Stockbridge may have been overturned and faulted prior to deposition of the Middle Ordovician pelite (Ratcliffe, 1965, 1968a). No exposures in this area, however, prove this, as overthrusting of the Walloomsac onto the inverted Stockbridge could also satisfy the geometric relations.

No mappable folds can be attributed to the pre-Walloomsac event, and the exact nature of this disturbance is unknown. However, high angle faulting accompanied by folding of consolidated carbonate rock seems to have been important.

D_1 Folds (Post-Walloomsac)

Isoclinal, highly deformed small folds (amplitude less than 100 feet) present in the Nassau, Everett, Walloomsac, and Stockbridge Formations are found to be refolded by two later fold systems corresponding to D_2 and D_3 events (Stop 6). These folds appear to lack an axial plane foliation and are most abundant closest to thrust contacts, suggesting that they may have formed during the emplacement of the allochthonous rocks.

D_2 Fold System

The prominent northeast structural trends were developed during an event that produced the regional well-developed axial plane foliation S_2. This foliation penetrates the autochthon-allochthon contact and thus post-dates the thrusting. Fine-grained lepidoblastic muscovite and chlorite lie in this foliation, giving the pelitic rocks a silky phyllitic sheen. The first metamorphism probably occurred during this event.

D_3 Fold System

A well-developed northwest-trending crenulation slip cleavage formed during a third phase of folding that affects the entire section. The intensity of the D_3 folding increases to the east and dominates the structural pattern in the Ashley Falls, Great Barrington, Stockbridge, and eastern part of the Egremont quadrangles. Locally in the western part of the State Line quadrangle a slip cleavage with a N. 50°E. strike is recognized along with the normal N. 15° to 20°W striking late cleavage.

METAMORPHISM

The grade of regional metamorphism increases progressively toward the east-southeast. Isograds of biotite, garnet, and staurolite in rocks of appropriate composition are shown on the maps (Figs. 4 and 5). The isograds trend N. 15° to 20°E. across the map.

Common mineral assemblages in the pelitic rocks are listed below in order of increasing rank (minerals in parentheses need not be present).

(1) Chlorite-sericite-quartz-albite-epidote-(stilpnomelane) (Ꞓev)
Chlorite-epidote-ilmenite-hornblende-actinolite-Na Plagioclase (Ꞓnv)
Chlorite-sericite-quartz (Ow)

(2) Chlorite-sericite-quartz-hematite-biotite (Ow)

(3) Chlorite-muscovite-quartz-ilmenite/magnetite-albite-paragonite-chloritoid (Ꞓev)

(4) Chlorite-muscovite-quartz-oligoclase-ilmenite/magnetite-chloritoid-garnet-(paragonite) (Ꞓev)
Chlorite-muscovite-biotite-quartz-plagioclase-garnet (Ow)

(5) Chlorite-muscovite-chloritoid-biotite-plagioclase-quartz (Ꞓev)

(6) Chlorite-muscovite-chloritoid-staurolite-garnet-plagioclase-quartz (Ꞓev)

Two generations of white mica and biotite are found; one lies in the regional foliation (S_2) and the other crosses S_2 or locally is alined in the late slip cleavage (S_3). Moreover, non-rotated garnet, staurolite, and chloritoid have grown across the slip cleavage and include crenulated S_2-oriented mineral fabric (largely sericite and ilmenite).

On the basis of these textural relations, the most intense metamorphism occurred after the development of the regional foliation (S_2) and was synchronous with or later than the development of the slip cleavage (S_3) (Ratcliffe, 1965, 1968a).

RELATIONSHIP OF METAMORPHISM AND TECTONIC DISTURBANCES

Evidently no metamorphism accompanied either the D_0 or D_1 events, but low grade metamorphism accompanied the formation of the D_2 fold system, producing lepidoblastic sericite, chlorite, and ilmenite.

Radiometric data (Zen and Hartshorn, 1966) indicate that the last phase of recrystallization may have been as recent as Devonian. The older metamorphism and foliation (S_2) may be Ordovician in age as suggested by the writer (1965, 1968a). This interpretation is consistent with the recent isotopic data from the northern Taconics (Harper, 1968) and with the observations of Ratcliffe (1968b) that the Cambrian and Ordovician rocks bordering the western edge of the Cortlandt Complex near Peekskill, New York, underwent regional deformation and metamorphism prior to intrusion of the igneous rocks thought to be at least as old as 435 m.y. (Long and Kulp, 1962).

The D3 event probably was Devonian in age. Although no evidence yet available permits correlation of these events with an absolute time scale, this scheme can be tested by future isotopic investigations.

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ROAD LOG

Take New York Thruway south from Albany, exit onto Berkshire Spur, and exit at interchange B-3 (Route 22) near State Line Massachusetts. Assembly point will be the Sunoco Station at the exit ramp from the east-bound lane of the Thruway at intersection of Rte. 22 and Berkshire Spur. Starting time will be 8:30 at this point. The drive from Albany should take approximately 50 minutes. Stops will all be in the State Line and Stockbridge quadrangles.

For early comers good exposures of the purple and green phyllites of the Nassau Formation (Ꞓnp) can be seen beside the Sunoco Station. Note multiple cleavages, folded lineations, and large pyrite crystals.

Road log starts from exit B-3 and Rte. 22 at the Sunoco Station.

<u>Mile</u>

0.0 Start Sunoco Station, intersection Rte. 22 and exit ramp of B-3 interchange, head north on Rte. 22, crossing Thruway (State Line quadrangle).

0.7 Turn left on Pleasant Valley Road (shown as Tunnel Road on topo. sheet).

1.6 <u>Stop 1.</u> Roadcuts in units e and d of the Stockbridge Formation beneath the Chatham slice. Folded folds (Fig. 7). Major structure in cut is a late (D_3) synform that contains unit d resting on unit e (inverted sequence). This synform folds an earlier lineation (bedding-cleavage intersection). Note minor faulting and calcite fracture filling in the D_3 structures. The map patterns in the carbonate rocks at this locality are more complex than those within the Chatham slice and in other areas of exposures of carbonate rocks. Rock above and below the thrust contains identical D_2 and D_3 structures, suggesting that the differences in map patterns probably are the result of the carbonates having been deformed severely prior to the development of the D_2 structures. Perhaps this early deformation took place at the sole of the Chatham slice during its emplacement.

Continue west on Pleasant Valley Road. Leave State Line quadrangle; enter Canaan, New York.

3.3 Turn left on Columbia County Rte. 5; cross Thruway.

4.0 Outcrops of subgraywacke interlayered with green and purple phyllites of the Nassau Formation.

4.4 Enter State Line; outcrops of purple phyllite with massive green subgraywacke dipping to south.

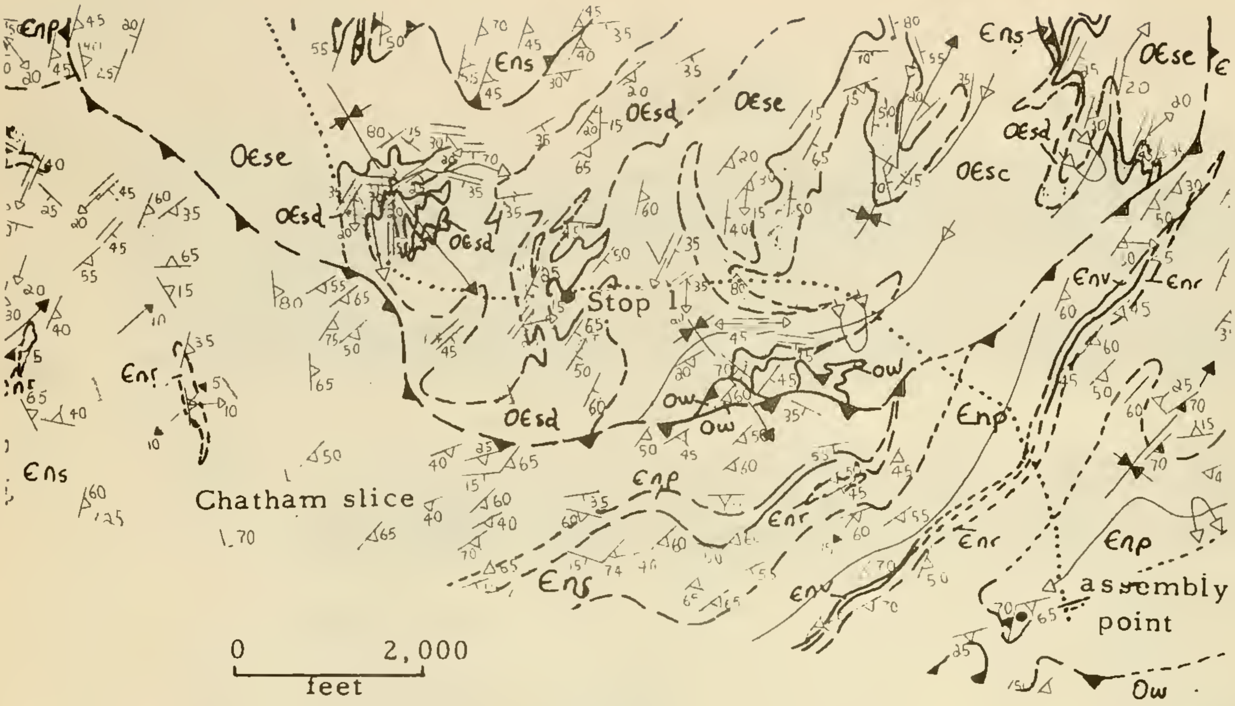

Fig. 7. Geologic map showing the locations of the assembly point and Stop 1, and contact relations between the autochthon and the Chatham slice. North-central part of the State Line quadrangle. Unit d of the Stockbridge is shaded.

Fig. 8. Geologic map showing locations of Stops 3 and 4, in the State Line quadrangle.

4.8 Stop 2. Nassau Formation of Chatham slice. Purple and green phyllite lower member (Ɛnp). Regional foliation S_2 is axial plane of small isocline that strikes N. 50°E. and dips 40°S.E. S_2 is crossed by a weakly developed northwest-trending cleavage (S_3) and a N. 10°E. 75°W. cleavage that is related to minor faults. Note small west-dipping reverse fault and chlorite and quartz mineralization in fractures.

Continue south on Columbia County Rte. 5.

5.3 School House Road on right.

5.4 Turn left on dirt road. Stop 3. (See Fig. 8 for location.) Metavolcanic rocks (Ɛnv) in Nassau Formation. We will hike up hill to volcanic rocks. First exposures in field will be of (Ɛns), the green colored phyllite and siltstone unit of the Upper Nassau. Massive graywacke forms the base of the cliffs and underlies the volcanic rocks. The contact with the metavolcanic rocks can be best located by watching for the absence of quartz. The lithology and interpretation are discussed in the text. Return to cars and continue south on Rte. 5.

5.8 Intersection Columbia County Rte. 24. Continue straight.

7.0 Turn left on Fog Hill Road. Outcrop of parautochthonous Walloomsac here resting on the Chatham slice.

7.4 Roadcuts of Ɛnv, Ɛns, and Ɛnr, same as Stop 3. If the road to No Bottom Pond is not passable, we will hike from this point to the Pond.

8.3 Intersection Fog Hill and Middle Roads; bear left and turn immediately onto dirt road to Mt. 10 ski club. Because this road is locally very muddy, only drivers having confidence in their superior driving skill should attempt this. Hopefully enough sure-footed vehicles will be available to make the drive successfully. Stay on the road! The ditches are deep and very soft.

9.3 Stop 4. No Bottom Pond window. Unit c of the Stockbridge unconformably overlain by limestones of the basal Walloomsac and exposed in a window. Surrounding rocks belong to Ɛns of the Chatham slice (see Fig. 8).

No Bottom Pond periodically empties via an underground drainage system, leaving the pond dry; hence it is aptly named. According to the owner, Professor Gordon of Columbia Teachers College, the water leaves the pond in the southwest corner where some large submarine caves can be seen. Evidently this carbonate rock is connected with the Stockbridge 600 feet lower in the valley to the north (Stop 1).

The structure is a doubly plunging anticline that exposes the window rocks. This indicates this is a true window rather than an isolated fault block.

Walk west and then south along marked trail; note outcrops of dolostone and solution collapse features.

a. Large cliffs of calcitic dolostone typical of unit c of the Stockbridge; fine laminations show the bedding strikes N. 35°W. and dips 20°S. W. Right-side-up cross laminations can be seen in some layers. Continue walking south along cliff to exposure of Walloomsac limestone (Owl).
b. Fossiliferous black limestone, containing crinoid, bryozoan, and algae debris overlies unconformably the Stockbridge. The limestone dips steeply southeast; bedding in the dolostone strikes N. 20°W. and dips 25°S. W. This angular discordance indicates the Stockbridge was rotated into a steeply dipping attitude prior to deposition of the Walloomsac.

The age of this unit has not been pinned down yet. The conodont assemblage studied by John Huddle, of the U. S. Geological Survey, indicates that the rock could be as old as the Early Rockland Formation of Central New York State. The fossil evidence suggests a middle to upper Ordovician age. Precise dating of this limestone would help date the mid-Ordovician unconformity as well as set a maximum age for the emplacement of the Chatham slice at this locality. Brachiopods and cup corals will be accepted!

Walk west across the dolostone to a small infold of Owl; continue to the Owl on the west flank of the window. The large boulders are blocks of Єnv traceable to Stop 3.

c. Continue west to small cliffs; exposures of Єns of the Chatham slice. Note bedding folds plunge southwest, away from the window. The foliation is S_2. (See Fig. 8.)

Return to cars.

Return to Fog Hill Road.

10.2 Turn right and then left onto Middle Road. Leave the Chatham slice and pass into a belt of Walloomsac that is thrust onto the Chatham slice.

10.7 Small outcrops of black Walloomsac and overlying greenish phyllite of the Everett slice.

11.2 Intersection with Fog Hill Road; continue straight.

11.8 Intersect Rte. 22; continue straight across highway on dirt road (Red Rock Road). Avoid right fork.

12.6 Turn right on paved road (West Center Road).

13.9 Stop 5. Everett slice, Everett Formation. Biotite zone. Walk up in pasture (beware of ram) to exposures of greenish phyllite and quartz phyllite of the Everett. Gritty, green quartzite layers define bedding that is folded with S_2 as axial plane. Folds plunge southeast. A weakly developed slip cleavage (S_3) is present, producing crinkles on S_2. The common mineral assemblage in the Everett at this grade is chlorite-chloritoid-albite-plagioclase-quartz (paragonite).

At west edge of pasture and on bluff are exposures of carbonate rock. Note the brecciation of the dolostone and the subsequent calcite vein filling. In the woods to the north these carbonates float in a matrix of Walloomsac. Blocks like this are found locally at the base of the Everett slice and have been interpreted as a tectonic breccia (Zen and Ratcliffe, 1966).

Return to car.

14.3 Bear right at intersection of Maple Hill Road.

15.2 Stop 6. Stockbridge unit g, Walloomsac, folded folds. Camera stop; no hammering please. Walk east up hill to exposures of calcite marbles with interbedded beige weathering dolostone characteristic of unit g of the Stockbridge. The contact with the overlying Walloomsac is exposed and evidently is conformable. Excellent exposures of folded folds, micro faults, and the various foliations S_2 and S_3 will be pointed out. D_1 folds that are cut by S_2 and S_3 can be seen at several localities.

Continue south on West Center Road.

15.9 Stop 7. Everett Formation, showing color banding and two cleavages, S_2 and S_3. Everett slice.

Dark-gray and lighter-greenish-gray color bands show the bedding that is cut by the foliation S_2 (N. 20°E. 50°S. E.). A late crenulation cleavage, locally a true slip cleavage, trends N. N. W. From this point eastward, the slip cleavage becomes well developed, and a late set of folds trending northwest-southeast deforms the earlier northeast-trending structures. Dark colored rocks such as these are found in the Everett, but more commonly lustrous green and light-greenish-gray colors predominate in the Formation as a whole. Abundant chloritoid can be seen in some of the lustrous magnetite-rich layers.

16.0 Continue east to intersection with East Street. Turn left.

16.8 Bend in road; swing right; outcrop unit c, Stockbridge.

17.7 Turn left on Rte. 41.

17.8 Turn right on Glendale Road; enter Stockbridge quadrangle (Fig. 5).

18.2 Stop 8. Stockbridge unit d overlain by Walloomsac and Everett. Area of strong D_3 folding.

This stop is located at the north end of a late D_3 north-plunging antiform (see Fig. 5). At the base of the cliffs, the axial surface of an isoclinal fold in sandy marble probably unit d trends east-west and dips north. This probably is a rotated D_2 fold.

Overlying this is an impure schistose calcite marble containing pebbles (?) of black limestone. This limestone marks the base of Walloomsac in this area. This marble grades up into sooty-black biotite-rich schist of the Walloomsac. Light-silvery-gray chloritoid-rich schists at the top of the hill are assigned to the Everett Formation. Mineral assemblages 4 and 5 are common here (see text).

Note the intense crumbling of the foliation in the schist units, and the west-dipping slip cleavage (S_3). This exposure points out the difficulty in distinguishing the Walloomsac and Everett at this grade.

19.2 Continue east on Glendale Road to intersection with Rte. 183. Turn left. Note exposures of Walloomsac showing D_3 folding. Everett exposed in this roadcut contains chloritoid and staurolite with no evidence of incompatibility.

21.6 Intersection Rte. 102. Turn right toward Stockbridge.

23.0 Intersection Main Street. Turn left. Continue through town of Stockbridge.

23.8 Intersection Rte. 7. Continue east on Rte. 102.

25.8 Turn left into parking lot of South Lee School at blinking light. Stop 9. Stockbridge unit c, basal Walloomsac (Owm). Light-gray crystalline calcite dolostones of unit c dip west under rusty weathering biotite schist and calcite schistose marbles of the Walloomsac (Owm) that are at least 250 feet thick. This facies of the Walloomsac is well developed in the eastern outcrop area of the carbonate belt, and appears to be related to areas of maximum erosion on the middle Ordovician unconformity.

Overlying this and interlayered with Owm are rusty weathering schistose rocks more typical of the Walloomsac. These schistose rocks can be seen in the pasture to the west. More aluminous schists of the Walloomsac are coarse garnet, staurolite, biotite-muscovite-quartz plagioclase schists at equivalent grades to the south in the Great Barrington quadrangle.

The structural style in the Stockbridge quadrangle is marked by isoclinally folded and refolded folds. The structural trends are strongly northwest-southeast, owing to pervasive development of the D_3 structures.

For Massachusetts Turnpike, follow Rte. 102 (turn left after leaving parking lot) and get exit 3 at Lee, Massachusetts, either west or east bound. For routes north or south, turn right on Rte. 102 and join Rte. 7 in Stockbridge.

Stratigraphy, structure, and metamorphism of the Taconic allochthon and surrounding autochthon in Bashbish Falls and Egremont quadrangles and adjacent areas[1]

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INTRODUCTION

The Taconic Range is underlain by pelitic schists varying in metamorphic grade from chlorite zone to staurolite-almandine zone. The stratigraphic position and structural history of these rocks have been part of the classical Taconic controversy. We now know that there are really two problems: 1) that of the "low Taconics," comprising the north end of the Taconic Range and all of the slate belt of Vermont and New York; and 2) that of the "high Taconics," comprising most of the Taconic Range from Dorset Mountain, Vermont, to at least Indian Mountain, Connecticut. In the low Taconics, the age of the rocks is known but the geometry is uncertain. Fossils indicate that the rocks are correlative with those of the surrounding Middlebury synclinorium and its southward continuation. In contrast, in the high Taconic region, the problem is one of known geometry but uncertain age. There has been general agreement, since the days of Dana (1884), that the schists are in synclinoria and rest above early Paleozoic carbonate units, but because these schists have never yielded fossils, assignments of their stratigraphic ages have always depended on long-range correlations, the credibility of which has depended on the bias of individual geologists.

Detailed bedrock geologic mapping in the southern part of the Taconic Range has been carried out since 1961 by the writer and by N. M. Ratcliffe. We have gathered data now to show that:

1. There is a mappable stratigraphy within the Stockbridge Formation in the country surrounding the Taconic Range. This stratigraphy is consistent with that of western Vermont (Doll et al., 1961) on one hand, and with those of the Pine Plains area (Knopf, 1962) and of the New York City area recently deciphered by Hall (1968) on the other. Fossils found within the highest subunit of the old Stockbridge Limestone, now reassigned to the overlying Walloomsac Formation, are of Middle Ordovician age; therefore the bulk of the Stockbridge Formation is now assigned with confidence to the Cambrian and Early Ordovician, in agreement with the age of comparable units in Vermont and New York, previously correlated on lithic basis alone.

[1]Publication authorized by the Director, U. S. Geological Survey.

2. An angular unconformity exists between the Stockbridge Formation and the overlying fossiliferous Middle Ordovician (probably Trenton) Walloomsac Formation.

3. The carbonate units have been deformed at least twice with noncoincident axes of folding.

4. The controversial pelitic schists of the Taconic Range (Everett Formation) occur as one or more sheets overlying the Stockbridge and Walloomsac Formations. At many places, the contact is marked by slivers of carbonates (in beds or blocks) of the Stockbridge Formation; by breccia of carbonate, Walloomsac, and Everett lithology, and by intimate interlamination of the black-grey schists of the Walloomsac and the green schist of the Everett. From this we conclude that the Everett Formation does indeed constitute an allochthonous mass, part of the complex of the Taconic allochthon. This local evidence is independent of regional correlations.

5. Available evidence suggests strongly that the rocks of the Taconic Range, especially the Everett Formation, have undergone two episodes of regional metamorphism, of which the later one presumably set the K/Ar and Rb/Sr isotopic clocks to an "Acadian" age (360 million years, more or less).

These new data and reinterpretations help to solve some of the problems of the high Taconic region but bring forth new problems which will be outlined later. Trips 3A and 3B show the outlines of the stratigraphy, within both the autochthon and the allochthon, demonstrate the main features of structure, particularly the interrelations of the autochthon and allochthon, and finally show some of the problems in metamorphism and related chronologic questions. Trip 3A will be concerned mainly with the stratigraphy and structure of the autochthon, whereas trip 3B will be concerned largely with the structure and metamorphism of the allochthon.

WARNINGS AND ACKNOWLEDGMENTS

To preserve the outcrops for future workers and visitors, hammers may not be used except where announced. Most of the features of interest can be seen, and seen better, on the naturally exposed and weathered surfaces. Several important bits of evidence have already been damaged by earlier visitors who hammered on the rocks before I had an opportunity to tell them of the significance of the features being bashed. I thank all participants of the trips, as well as future users of this guidebook, in advance for their cooperation. I also thank the landowners for permission to visit the outcrops. The Egremont quadrangle was mapped jointly with Nicholas M. Ratcliffe, and I am indebted to him for allowing me to present the geological map of the entire quadrangle before its publication. Ratcliffe, John S. Dickey, Jr., and Nicholas Lampiris worked with me at various times in the field, and I thank them for their able assistance. D. S. Harwood, N. M. Ratcliffe, J. C. Reed, and R.S.Stanley commented on the manuscript and on the outcrops; they suggested many improvements. Responsibility for the data and interpretations, however, are mine alone. I thank my lucky stars for escape from serious injury despite extensive field use of defective automobiles.

STRATIGRAPHY

The lithostratigraphic succession and lithic characteristics and thicknesses of units in the field-trip area are given in table 1 (p. 12). A generalized geologic map of the Egremont and Bashbish Falls quadrangles and parts of adjacent quadrangles is given in figure 2 (see fig. 1 for location).

Fossils have been found in a thin-bedded blue-grey limestone near Pittsfield, Mass., at a stratigraphic position corresponding to the basal unit of the Walloomsac Formation (see Zen and Hartshorn, 1966, text). These fossils include cup corals that are no older than Black River and more probably Trenton in age. Fragmentary brachiopods and hash of pelmatozoan stems in a basal thin-bedded, phyllitic black limestone of the Walloomsac Formation (stop A-7) are of similar age. Thus, the rock units here assigned to the Dalton, Cheshire, and Stockbridge formations are most likely Paleozoic but pre-Trenton. The lithostratigraphic equations given in table 1 thus seem probable, though these equations do not demand the suggested correlations in detail.

Subunits d and f of the Stockbridge Formation have proved particularly troublesome to map. Unit d separates a massive, grey dolostone from a massive pure-white to grey-mottled calcite marble. Unit d is locally only a few feet (1 meter or more) thick but elsewhere, particularly in the north-central part of the Egremont quadrangle (fig. 2), a large assemblage of heterogeneous units is mapped as part of unit d because of the stratigraphic position of the units. As mapped, unit d must be lenticular and interdigitate laterally with units c and/or e; interdigitation on scale of a few tens of feet is locally visible. Similar variability and uncertainty exists for the contact of units f and g, which closely correspond lithically to units d and e. We will see this variation at stops A2 and A3.

The base of the Walloomsac Formation, which unconformably overlies the Stockbridge Formation, is at places a black phyllite (stop B6), at places a thin-bedded silty-micaceous black-blue limestone (stop A7), and at places a massive ferruginous silty limestone, typically weathering brown, and as much as 200 feet (70 meters) thick. This last lithic type is best exposed on the east shore of Lake Washining. Along this basal zone, both in the trip area and beyond, iron ores were mined in the 19th century. This iron-rich material may represent weathering residua accumulated on the unconformity surface and redeposited with the younger sequence. Where both residua and carbonate deposits are absent, typical black phyllite of the Walloomsac directly overlies the Stockbridge Formation.

A large area of black phyllite occurs in the northwestern part of the Bashbish Falls quadrangle and adjacent part of the Egremont quadrangle (stop B2). This area is completely enclosed within the Everett Formation (fig. 2). On the published geologic map of the Bashbish Falls quadrangle (Zen and Hartshorn, 1966), this unit was mapped as the Egremont Phyllite. The rocks are very similar to those of the Walloomsac Formation and occupy a corresponding geometric position. In fact, east of Mount Fray (Egremont quadrangle), the Walloomsac and the Egremont are separated only by a high ridge and occur at about the same elevation on the two flanks. On figure 2, therefore, the area is shown as underlain by the Walloomsac Formation, even though carbonate slivers, so typical of the Walloomsac-Everett contact over much of the trip area, are absent here. The structural implication of the Egremont-Walloomsac equation is discussed in the section on structures.

Within the allochthon, massive quartz (rarely feldspar) pebble conglomerate, having a green silty phyllite matrix, and interlaminated with quartzite and phyllite on 1/8 inch (3 mm) scale, occurs locally in the Everett Formation in the western part of the trip area (fig. 4). At Cedar Mountain near Bashbish Falls, it is at least several hundred feet (100 meters or more) thick. This unit is here lithically equated with the Lower Cambrian(?) Rensselaer Greywacke of the low Taconic sequence (Zen, 1967). The conglomerate sequence is much thinner than the Rensselaer of the Rensselaer Plateau area, but this is in accord with the previous suggestion (see Zen, 1967, p. 49) that the high Taconic sequence is the eastern facies equivalent of the Rensselaer. The Rensselaer is a turbidite deposit (Bird, 1963) which had a generally westerly source, so the restriction of the massive conglomerate to the western periphery of the high Taconic sequence makes sense. If the correlation of the conglomerate in the Everett with the Rensselaer and, by implication, the Pinnacle Greywacke of the Camels Hump Group of Vermont (see Zen, 1967; Doll et al., 1961) is correct, then the bulk of the Everett would be correlated with the Underhill Formation of northern Vermont (Doll et al., 1961) and with the Hoosac and possibly part of the Pinney Hollow Formation of southeastern Vermont (Doll et al., 1961). This is in agreement with the previously suggested correlation (Zen, 1967) made before these conglomerate bodies were recognized in the trip area.

Rocks of the autochthonous Dalton, Chesire, and Stockbridge Formations were laid down in a shelf environment where the water was shallow and where many reefs were probably being built. To the east, the shelf terminated abruptly against the deep basin, part of the eugeosyncline (see Rodgers, 1968; Zen, 1967, fig. 14). In Early and Middle Ordovician time, downfaulting of blocks of the shelf, related to upfaulting in the eugeosyncline, led to an overall deepening of water on parts of the shelf and to local derivation of clastic sediments from the eugeosynclinal area, accounting for the unconformity at the base of the Walloomsac Formation and for the change in sedimentary regime at Walloomsac time (Zen, 1967). Locally, however, the shelf was raised above sea level and became eroded and weathered, as indicated by the iron-rich basal Walloomsac Formation.

Within the deep basin itself, deposition of the Everett Formation started earlier than the beginning of Dalton sedimentation and had a western cratonal source before the evolution of the shelf cut off sediment supply from that direction. If the regional stratigraphic correlations suggested by the writer in 1967 (plate 2; fig. 5) are correct, this cutoff was approximately synchronous with the deposition of the uppermost exposed Everett Formation, though lack of stratigraphic markers in the Everett makes any more detailed assignment meaningless. Nonetheless, sedimentation within the deep trough presumably continued without a major break, though the direction of sediment transport may have shifted with the rearranged hydrologic pattern after the establishment of the shelf, and, very likely, the ultimate source of sediments probably also shifted from a western craton to eastern eugeosynclinal islands. Rocks of the upper Bull Formation or younger sediments (Zen, 1967, fig. 5), which are lithically distinctive, are not present in the Everett Formation of the trip area (with the possible exception of one small area in northwest Egremont quadrangle, in a structurally separate unit; see section on structure).

STRUCTURE

Structures within the autochthon

Within the autochthon, two overturned to recumbent folds extend en echelon from the northern part of the Egremont quadrangle to the southern part of the Bashbish Falls quadrangle. The lower, more southern of these folds has been designated the Foley fold (Zen and Hartshorn, 1966). The axial trend of the folds is much distorted by later folding, but the average trends are nearly north. The recumbent nature of the folds is brought out by later refolding along more northwesterly axes, which crenulated the contacts between units, for example unit c of the Stockbridge and higher members of that formation, and led to the presence of younger units within cores of anticlines which plunge south-southeast in response to the geometry of the early folding (stop A4).

West of the Taconic Range, the Stockbridge Formation has been folded, but the present mapping is insufficient to provide a detailed picture. The carbonates have clearly been faulted by a series of high-angle faults, but these faults may be the latest event (see below).

Whether the Walloomsac Formation, which unconformably overlies the Stockbridge, partakes in the early recumbent folding as indicated in fig. 2, is not clear. Near Alford, Mass., isolated patches of Walloomsac rest on overturned units of the Stockbridge; the relation could be construed as evidence of pre-Walloomsac folding of the Stockbridge, but the Walloomsac might have been thrusted in after the folding.

Structure within the allochthon

Although the overall map pattern (fig. 2) shows the allochthonous rocks in the trip area in a single structural sheet, the sheet very likely is a composite of several thrust slices. Lack of units within the Everett Formation that can be traced for more than a few thousand feet, however, leaves the internal structure of the allochthon uncertain. Where traceable beds do occur locally, the rocks are seen to be intricately folded on a small scale.

We do know, nonetheless, that rocks of the allochthon have been deformed at least twice. Evidence for this is the presence of folded foliations (stop B1, B2, B4, B5). The trends of the two foliations (the latter commonly a slip cleavage) broadly agree with the early and late fold trends in the Stockbridge Formation. A possible interpretation of these data will be given later.

Structures along the base of the allochthon

The contact between the autochthon and allochthon is marked by a zone of intense tectonic movement and locally inclusion of exotic rock masses (stop A6). Most readily recognized among these masses are slivers of carbonates, which can be identified as units of the Stockbridge (Zen and Ratcliffe, 1966). Within the slivers, the carbonates can be recognized locally as beds, but elsewhere they are blocks in a breccia associated with rarer blocks of phyllite, identified as the Walloomsac and the Everett formations. The limestones appear to have flowed whereas the dolostones shattered and were later healed with quartz and calcite (fig. 3). Associated with these breccias and slivers are tectonic interlaminations of black and green phyllite. Both the early and late cleavages are recognized in the movement zone. Because the tectonic contact zone is mapped on both the east and west sides of belts of outcrop of the Everett, as well as at the plunging ends of such belts, the slivers are not the product of local upthrusting, and the relations constitute independent structural evidence for the allochthon.

On the west side of the Taconic Range, a repetition of allochthonous rocks below some Walloomsac shows that the Walloomsac itself is locally at least parallochthonous. One patch of rocks mapped as the Everett, about 1 mile (1.6 km) south of stop A6, may be the remnant of an older slice of the Taconic allochthon (Zen, 1967). Small areas of isolated Everett lithology within the authochthon, not shown in fig. 2, are interpreted as debris of the allochthon left behind as it moved westward. If so, then the present erosion surface must closely approximate the surface over which the allochthon moved, thereby making it possible for the moving allochthon to pick up pieces of the Stockbridge, to be preserved as slivers.

Just north of Route 23 (fig. 1), the Taconic Range is actually mainly underlain by the Walloomsac Formation, the Everett being preserved only in a slender syncline on the east flank (stop B1). Thus even though the Taconic Range on the whole is a synclinorium, locally, as in the trip area, a late anticline is developed along the central part of the synclinorium. This is significant because it makes more reasonable the suggestion that the Egremont Phyllite and the Walloomsac Formation are the same unit, and furnishes means for estimating the preserved thickness of the allochthon. This thickness cannot be more than a few thousand feet and is the basis for the tentative estimate (table 1) of the thickness of the Everett itself.

The Everett and Walloomsac have been faulted together by at least one high-angle fault having a pattern similar to that involving the authochthonous Stockbridge Formation and Walloomsac Formation west of the Taconic Range. These high-angle faults may well record the last significant diastrophic event in the area.

METAMORPHISM

As indicated in table 1, rocks of the trip area have been metamorphosed over a wide range of grades. In the northern half of the area (see fig. 1), the Everett and Walloomsac are typically fine-grained chlorite-muscovite-quartz-albite phyllites. Chloritoid is found in phyllites near Alford; almandine is found in the Long Pond area, in the Jug Eng-Mount Bushnell ridge area, and thence southwest across the Bashbish Falls quadrangle. Staurolite is found approximately southeast of a line trending southwest from Mt. Race. Kyanite, though reported by Agar (1932, p. 38), has not been verified by the present work. Paragonite is found sporadically in the Everett Formation. The pairs chloritoid-biotite and chloritoid-staurolite are common, and cummingtonite-almandine-biotite rocks are locally traceable within the higher grade Walloomsac near Lions Head. Both albite and paragonite (rarely both) are found with chloritoid. Epidote is ubiquitous within the small-pebble conglomerate, here equated with the Rensselaer; it is definitely not a detrital mineral in these rocks.

Within the carbonate units, phlogopite is the most common iron-magnesium silicate (stop A4). Hornblende is found only in calcareous beds of the Walloomsac in the southeastern part of the trip area, where the metamorphic grade is the highest. Unit a of the Stockbridge characteristically contains white radiating clusters of tremolite near the eastern margin of the trip area, in beds originally rich in quartz, but its appearance is apparently not related to faulting as interpreted by Hobbs (1893b, p. 794). This mineral is also found at the contact between the Walloomsac and unit c of the Stockbridge (stop B6).

Petrographic and electron microprobe studies show that the plagioclase in the Everett Formation, over a broad belt that trends northeast across the northern Bashbish Falls and southern Egremont quadrangles, is zoned in a peculiar way: the core grades outward from a high-Ca albite into an oligoclase, but then a sharp contact separates this zone from an outermost zone of nearly Ca-free albite. The gradational zoning outward into a calcic feldspar is what one would expect in prograde regional metamorphism, but the sharp reversal in zoning must be associated with a pervasive regional retrograde metamorphism. In some feldspar grains, the high-Ca zone is completely altered to a micaceous mixture, the mica foliation being parallel to the foliation in the matrix of the rock.

Regional metamorphic grade generally increases towards the southeast in this part of New England (see Thompson and Norton, 1968). The higher grade rocks have yielded Acadian K/Ar and Rb/Sr isotopic ages (see Long, 1962; Zen and Hartshorn, 1966; Clark and Kulp, 1968), whereas the low-grade rocks in the northern part of the Taconic allochthon have yielded older ages (Harper, 1968). It is therefore suggested that the higher grade metamorphism in the southern part of the trip area is Acadian, whereas the low-grade metamorphism in the northern part is an earlier event. The Acadian metamorphic grade rises steeply to the south. The belt of peculiar feldspar (stop B2) represents the first effect of the Acadian event, detectable today only because the plagioclase is relatively coarse grained. Any early effect originally present in the Acadian high-grade zone has been wiped out; the higher grade rocks show no reversals in the zoning of the feldspars, and one of these samples (stop B5) has been dated as 355 m.y. on a biotite (Rb/Sr) and 390 m.y. on a coexisting muscovite (K/Ar). The conformity of mica folia in the retrograded feldspars to the foliation in the matrix, mentioned earlier, lends support to the idea that the fabric even in the retrograded zone has been so reorganized by the later metamorphic event that the isotopic clock was completely reset.

The implications of a pre-Acadian regional metamorphism are far reaching; we are currently making a systematic study of isotopic ages in the Taconic rocks with this possibility in mind. The following section presents a consistent chronology of events on the assumption of two metamorphic events.

SOME PROBLEMS IN CHRONOLOGY

The preceding summary shows that the following large-scale geologic events occurred in the area: (1) deposition of the miogeosynclinal rocks; (]) simultaneous deposition, presumably to the east, of eugeosynclinal sediments of the Everett Formation; (3) two folding events in the carbonates, the earlier involving recumbent folds; (4) at least two deformations in the Everett Formation; (5) development of a regional unconformity between the Stockbridge and Walloomsac Formations; (6) emplacement of slices of the Taconic allochthon; (7) two regional metamorphic episodes, the earlier one of low grade and the later one of rapidly rising grade to the southeast; (8) development of high-angle faults.

Sedimentation, with miogeosynclinal deposits on the platform to the west and eugeosynclinal deposits in the basin to the east, proceeded without large-scale interruptions from the earliest Paleozoic until about the end of Early Ordovician time (Zen, 1967). In Middle Ordovician time, in response to tectonism to the east that eventually led to the emplacement of the early slices of the Taconic allochthon (Sunset Lake and Giddings Brook slices; Zen, 1967), the sedimentary regime of the platform changed from carbonate deposition to black shale deposition. Concomitant tectonic activities in the miogeosyncline led to the development of an angular unconformity at the base of the black shale. Although this phase of tectonism may have included the early, recumbent folding in the carbonates, the tectonic style of the recumbent folding suggests a deeper seated environment. Instead, the writer now suggests that the recumbent folding accompanied the later emplacement of the allochthonous rocks of the high Taconic sequence (part of the "Dorset Mountain Slice"; Zen, 1967) because penetrative shearing, breccias, recumbent folds, and tectonic intercalation of rock units along the contact show that the deformation was not due to soft rock sliding and must have occurred in indurated rocks, perhaps under significant load. The patches of Everett, interpreted as remnants of thrust sheets, within the areas of Stockbridge Formation show that the surface over which the allochthon moved was approximately the present erosion surface, so that the development of the recumbent folds by interaction with the allochthon seems reasonable. This sequence of events is consistent with the relative ages of unconformity and recumbent folding shown at stop B6.

The age of the earlier metamorphism relative to the emplacement of the allochthon cannot be fixed until we can show either (1) that the autochthonous rocks have similarly undergone two metamorphic events of comparable grades, (2) that the earlier metamorphic fabric in the Everett was affected by a fabric related to the emplacement of the slices, or (3) that a metamorphic discordance exists between the allochthon and autochthon at least over part of the area of the high Taconic region. So far, all we can be sure of is that both the earlier metamorphic event and the arrival of the Everett as part of the Taconic allochthon predated the peak of the Acadian thermal event and postdated the emplacement of the Giddings Brook slice of the Taconic allochthon, a late Trenton event. Furthermore, progressive albeit slight change of metamorphic grade in the Taconic allochthon, without regard to boundaries of individual slices, makes it plausible that the metamorphism occurred after all the slices had been emplaced. Isotopic geochronologic studies, now initiated in the area by Marvin Lanphere and the writer, should shed light on this important question and allow a better synthesis of data from the trip area with regional relations for the northern Appalachian belt.

Because of the similarity in the structural orientation, the early deformation of the Everett Formation and of the Stockbridge may well have been simultaneous. The considerations of the preceding paragraph would then suggest that the early deformation of the Everett was associated with the emplacement of the slice.

The late deformation in the Stockbridge, the Walloomsac, and the Everett certainly occurred when these rocks were already in their present relative position, except for possible minor adjustments. The orientations of the structural features in these units seem to broadly agree; these features presumably define the Acadian structural grain of the region. The Acadian metamorphic trend, however, for unknown reasons does not parallel the structural trend, but cuts the later trend at a relatively high angle (see also Clark and Kulp, 1968). The metamorphic trend nearly parallels the present trend of the structural front of the Hudson and Housatonic massifs; one is tempted to speculate on a causal relation between them. But the structural history of these massifs is itself an enigma wrapped in mystery, and so we must leave this discordance between late structural and late metamorphic trends as a phantom mystery of northern Appalachian knowledge.

A few final words should be added about the high-angle faults in the area. Near Alford in the northeast part of the Egremont quadrangle, Zen and Ratcliffe (in press) have shown that the map pattern requires a hidden pre-Walloomsac fault in the Stockbridge Formation, which they supposed to be a high-angle fault. Such high-angle faults are well known in the Middlebury synclinorium (see Zen, 1968, for summary); similarly Thompson (1967, p. 87) has shown the need for a concealed pre-Walloomsac fault in western Vermont. Faulting of this age thus seems definite.

The high-angle-faulted Everett-Walloomsac contact in the southwestern part of the Egremont quadrangle indicates similar deformation after the arrival of the allochthon, a conclusion that agrees with mapping in the northern part of the Rensselaer Plateau (Potter, 1963). The straight trace of these later high-angle faults, clearly shown by topographic lineaments in the Hillsdale quadrangle (fig. 2), suggests that these faults may be the last major structural event in the area, postdating the Acadian folding. Acadian dislocations may account for the thrusting of Cheshire over the Everett in the northeastern part of the Egremont quadrangle, for the faulting west of Vossburg Hill, and for similar small-scale faulting in the autochthon.

Table 1.--Stratigraphic Data

Bedrock of the allochthon

Everett Formation: (Єev on fig. 5) Probably in excess of 2,000 feet (700 meters). Mainly quartzose and less commonly micaceous phyllite in shades of green and grey green; locally purple. Included in the unit are massive small-pebble conglomerate layers that grade into and is interbedded with impure quartzite layers; the pebbles consist predominantly of quartz with subsidiary feldspar and granite; matrix consists of granulated quartz, plagioclase, chlorite, muscovite, with rarer, disseminated epidote, stilpnomelane, and magnetite. In the northern part of the trip area, the Everett Formation is a chlorite-albite phyllite, occasionally containing chloritoid, but the metamorphic grade rises rapidly so that it is an almandine-chloritoid-chlorite schist near Mount Bushnell and an almandine-chlorite-staurolite schist east of Mount Everett. The unit is here equated lithically with the Greylock Schist of Mount Greylock, with the Mount Anthony Formation of MacFadyen (1956) of the Equinox and Bennington quadrangles, Vermont (Hewitt, 1961; MacFadyen, 1956), with the Hoosac and part of the Pinney Hollow Formations of southern Vermont, and with the Underhill Formation, including the Fairfield Pond Member of Doll et al.(1961) of northern Vermont.

Carbonate slivers and tectonic breccia: (OЄss and OЄt on fig. 5) Dominantly carbonate, as beds, blocks, or chips, ranging from fraction of inch (1 cm) to several feet (2-3 meters) across. Lithically the pieces and layers in the tectonic breccia can be assigned to the Stockbridge Formation, the Everett Formation, or the Walloomsac Formation.

Bedrock of the autochthon

Walloomsac Formation, including the Egremont Phyllite (Ow and Oeg on fig. 5): Probably about 1,500 feet (500 meters) thick. Jet-black to silvery-grey, fine-grained, silty carbonaceous, calcareous or pyritiferous phyllite or slate. Locally at or near the base of the unit are white or blue-grey limestone beds as much as several feet (1 meter) thick. In the southeast corner of the Bashbish Falls quadrangle and adjacent areas, the basal unit is locally a massive cliff-forming grey schistose marble, as much as 200 feet (70 meters) thick, mottled by dark irregular 1/2-inch (1 cm) phyllite blebs. The formation is a muscovite-chlorite-stilpnomelane phyllite over most of the area, but in the Bashbish Falls quadrangle its metamorphic grade rises southeastward, so the rock becomes an almandine-staurolite-biotite-oligoclase schist near Lions Head. Local layers of almandine-cummingtonite-biotite-magnetite-quartz rock. The Walloomsac is Middle Ordovician, in part because of fossils of that age in the basal beds; it is correlated with the Ira Formation (Zen, 1964a; see also Doll et al., 1961) and with the Hawley Formation and possibly part of the Moretown Formation of west-central Massachusetts (Hatch, 1967).

Stockbridge Formation: Total thickness estimated at about 3,500 feet (more than 1,000 meters). Unit g (O€sg on fig. 5): Massive, white to grey, mottled calcitic marble or limestone; interbeds of uniform, compact, homogeneous, pale yellow-grey dolostone a meter thick or less, and local interbeds of thin-bedded silty marble and phyllite. Thickness about 400 feet (130 meters). Correlated with the Chipman Formation and part of the Bascom Formation of Vermont (Doll et al., 1961), and the Copake Limestone and part of the Rochdale Limestone of the Pine Plains area (Knopf, 1962).

Unit f (O€sf on fig. 5): Discontinuous, heterogeneous unit that includes tan-weathering calcareous sandstone, commonly cross-stratified; grey, fine-grained silty limestone; massive calcareous sandstone; massive grey dolostone; grey friable dolostone; silty, dark-grey calcareous slate. Thickness abruptly varies along strike and ranges from 0 to 300 feet (100 meters). Correlated with the Cutting Dolomite of Vermont, but the thicker, more heterogeneous parts may include rocks mapped in Vermont as the Bascom Formation, which is also a heterogeneous unit including comparable rock types (see Doll et al., 1961).

Unit e (O€se on fig. 5): Very similar to unit g above, but locally the basal beds are a massive dolostone. About 400 feet (130 meters) thick. Unit e is here correlated with the Shelburne Formation as recently revised by Thompson (1967), or the Columbia Marble Member of the Shelburne Formation as used by, for example Doll et al. (1961) and Zen (1964b). It is also correlated with at least part of the Halcyon Lake Formation of the Pine Plains area (Knopf, 1962).

Unit d (O€sd on fig. 5): Very similar in lateral variability and in lithic assemblage to unit f, but includes a grey limestone-matrix, dolostone and limestone pebble conglomerate. Thickness estimated at between 0 and 250 feet (80 meters). Unit d is correlated with the higher members of the Clarendon Springs Dolomite as used by Thompson (1967), or the lower members (Sutherland Falls Marble and intermediate dolomite) of the Shelburne Formation as used by Doll et al., (1961); also with parts of the Halcyon Lake Formation of the Pine Plains area (Knopf, 1962).

Unit c (O€sc on fig. 5): Massive to thin-laminated, pale grey-weathering, white to iron-grey dolostone; in wooded areas outcrops are commonly low, rounded blocks or solution-widened pavement exposures. Individual beds as much as 5 meters thick. Stratification on millimeter scale, defined by concentrations of floating rounded quartz sand grains, locally shows cross-stratification or sedimentary slump features. Impure dark-grey chert beds as much as 1 foot (30 cm) thick occur in the western areas; these beds metamorphose into quartz nodules and quartzite-looking layers. Thickness about 700 feet (230 meters). Correlated with the Clarendon Springs and Danby Formations of western Vermont (Doll et al., 1961), and with the Briarcliff Dolomite of the Pine Plains area (Knopf, 1962).

Unit b (O€sb on fig. 5): Predominantly dolostone, but includes numerous black or silver-grey silty to phyllitic seams and subsidiary rusty-weathering quartzite or arkose beds. In most of the Egremont quadrangle and adjacent parts of the Hillsdale quadrangle, it can be further divided into three units: the highest unit, b3, is a brown- to tan-weathering grey, salmon, pink, yellow and white compact dolostone, having interbeds of grey silty to green micaceous phyllite, calcareous siltstone, rotten-weathering dolomitic sandstone, and calcareous and feldspathic sandstone. The intermediate unit, b2, is a grey-weathering, dark-grey massive dolostone having but minor phyllite partings. The lowest unit, b1, is much like the highest unit but contains fewer silty and sandy beds. Unit b is estimated about 600 feet (200 meters) thick. It is equated with the Winooski Dolomite and at least part of the Monkton Formation of Vermont (Doll et al., 1961), and with the Pine Plains Formation and possibly the upper part of the Stissing Dolomite of the Pine Plains area (Knopf, 1962).

Unit a (O€sa on fig. 5): A pale-grey, white, and grey-and-white-mottled, massive, grey- or white weathering dolostone. Near its base the unit contains vitreous quartzite beds half-inch to 10 feet (1 cm to 3 meters) thick. Rosettes of tremolite characterize the lower part of this unit; their occurrence apparently is controlled by the presence of quartzite beds in areas of higher metamorphic grade. The unit is about 700 feet (230 meters) thick. The unit is equated with the Dunham Dolomite of Vermont (Doll et al., 1961), and with the bulk of the Stissing Dolomite of the Pine Plains area (Knopf, 1962).

<u>Cheshire Quartzite</u> (€c on fig. 5): More than 300 feet (100 meters) thick. The unit is a pale-buff, white, grey, vitreous, massive quartzite, weathering glistening white. Bedding is rarely visible. Top of the Cheshire grades into unit a of the Stockbridge by acquisition of dolomitic cement, which helps to reveal thin bedding-lamination, and by interbedding of massive quartzite and dolostone. The unit is lithically equated with the Cheshire of Vermont and northwestern Massachusetts (Doll et al., 1961) and with the Poughquag Quartzite of the Pine Plains area (Knopf, 1962).

<u>Dalton Formation</u> (€d on fig. 5): Exposure of the unit is extremely limited in the trip area; thickness, estimated from adjacent areas, is about 200 feet (70 meters). The rock ranges from a massive to thin-bedded quartz-feldspar-biotite-muscovite-tourmaline granulite to biotite schist and to impure quartzite, all of which weather dull grey to orange-brown. The unit grades upward into the Cheshire by increase of quartzite beds. The unit is equated with units mapped as the Dalton in Vermont (Doll et al., 1961).

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ROADLOG TRIP 3A

Mileage

0 Start of trip at A & P store's parking lot, Great Barrington, Mass. Head south on Route 23.

0.3 Bear right with Routes 23 and 41 at traffic light.

1.3 Cross Green River; view of Housatonic valley to the left.

1.5 Turn left at sign "Wyantenuck Country Club."

2.1 Wyantenuck Country Club to the right.

2.7 Stop A1.

Dalton Formation; Cheshire Quartzite; basal unit a of the Stockbridge Formation. The rocks along the road are biotite-feldspar-quartz granulites assigned to the Dalton Formation, the lowest unit exposed in the trip area. Go west up hill. The Dalton is succeeded by a vitreous, massive, buff quartzite, the Cheshire; the contact appears abrupt. About 300 feet (100 m) of the Cheshire is exposed. Unit a of the overlying Stockbridge Formation is first encountered on the steep slope; along the contact is a single outcrop of a feldspathic biotite phyllite. Go up the steep slope, which is the dip slope of the overturned unit a, here a massive white dolostone. Half a mile north along strike, behind the club house of the Wyantenuck Golf Club, however, unit a is a mottled grey dolostone; these outcrops are now concealed "for aesthetic reasons."

The top of Vossburg Hill is underlain by unit a. Also present is a vitreous quartzite, about 15 feet (5 m) thick and traceable for more than 1,000 feet, nearly identical with the Cheshire but interbedded with unit a. Locally the quartzite shows inch-thick bedding visible on deeply weathered surfaces, apparently caused by minute amounts of carbonate. Thus the Cheshire-unit a contact is gradational in two ways: The two rock types interbed, and the quartzite in the contact zone is more calcareous than normal.

Tremolite crystals as much as 2 inches (5 cm) long are found in this stratigraphic position in the trip area, including the south end of Vossburg Hill.

Return directly to the car, proceed to mileage 3.0 to turn around and proceed north.

Mileage

4.5 Same corner as mileage 1.5; turn left on Route 23.

4.9 Intersection with Route 71; continue on Routes 23 and 41.

6.2 Mount Everett Country Club; view of Mt. Everett.

6.6 Bear right with main road at village of S. Egremont.

7.3 Junction of Routes 23 and 41; bear left with Route 41 south. Rocky ridge ahead is the Jug End-Mt. Bushnell ridge, underlain by the allochthonous Everett Formation. Sharp break in slope seen on right end of the ridge is the contact with the underlying autochthonous carbonate rocks.
Caution: Dangerous intersection

7.8 Road cut along Route 41. Stop A2.

Units c through g of the Stockbridge Formation. The roadcut on both sides of the road is grey, massive-bedded dolostone, the top beds of unit c of the Stockbridge Formation. East of the road at waist level is a thin-bedded silty limestone at the base of the heterogeneous unit d; the contact of c and d is concealed. Go east up the embankment. In the first clearing, rusty-weathering dolostone interbedded with brown-weathering calcareous sandstone is the main part of unit d. The next clearing across shows grey-mottled massive calcite marble, unit e. Across a gap, the next outcrop east at the low ledge is a second unit of brown-weathering calcareous sandstone, showing possible cross-stratification indicating an upright section. This is unit f. Continue uphill; the highest clearing is underlain by massive white and grey marble, locally interbedded with pale-buff dolostone (much boudinaged) that weathers into a "thread-scored beeswax" texture; the marble and dolostone are unit g, here in the core of a complex refolded syncline.

Although data from this traverse, from d to g, could be interpreted as a structural repetition of just two lithic units, regional mapping shows this is not so, and the lithic repetition is stratigraphic. At places, for instance west of Berkshire School in the Bashbish Falls quadrangle, outcrops of d and f in a single section show crossbedding that bears out this conclusion.

AT THE REQUEST OF THE LANDOWNER, THERE WILL BE NO SMOKING ON OR NEAR THE STOP. THERE IS A LOVELY 1753 HOUSE ON THE LAND AND ANY BRUSH FIRE COULD LEAD TO DISASTER. THANK YOU.

Proceed south to turn around.

Mileage

8.0 Turn around at side road; proceed north on Route 41.

8.5 Turn left at sign "Jug End Resort" onto secondary paved road.

9.3 Turn left onto tertiary paved road, "Avenue Road."

9.9 Dirt road junction. Bear left.

10.3 Stop A3; park at picnic area provided by the Isaak Walton League. Go west into ravine behind picnic table; beware of poison ivy.

Lithic variations of unit f and some problems in correlation. Outcrop in the ravine next to the road (elevation 840 feet) is marble of unit e. At the 880-foot level in the ravine is another outcrop of unit e. From the 920-foot level on, begin heterogeneous carbonate units, including: Massive grey calcitic dolostone and interbedded white marble; massive white dolostone; thin-bedded grey marble; calcareous sandstone; calcareous siltstone; silty grey limestone. These units are locally folded together. Where the stream branches, follow the northern branch. The section is capped at the 1,000-foot level by a black phyllite of the Walloomsac Formation; however, the Walloomsac-Stockbridge contact here transects contacts of units e, f, and g, and so presumably marks an unconformity or a fault. Continue up streambed and at the 1,020-foot level return to the Stockbridge Formation, here the massive grey and white marble of unit g. Marble continues to the base of steep slope (elevation 1,140 feet) which marks the contact with the Everett Formation (plagioclase-chloritoid-chlorite-almandine schist).

The thick, heterogeneous sequence of carbonate rocks in this traverse, all assigned to unit f of the Stockbridge, is in sharp contrast with the same unit at stop A2. The writer interprets the relations as a lateral facies change between the massive marble of parts of units e and g at stop A2 and the heterogeneous carbonate rocks of this stop. The presence of calcareous sandstone typical of unit f of stop A2 in this traverse supports this idea.

Continue on gravel road (Curtis Road south.)

11.7 Return to Route 41; bear right, proceed south on Route 41.

12.1 Road cut is in unit c of the Stockbridge Formation.

12.2 Spurr Pond on the left.

12.7 Berkshire School on the right.

13.2 Bear left on Berkshire School Road.

13.4 Stop A4 just beyond bridge.

Core of Foley Fold; inverted sequence of units c, d, e of the Stockbridge Formation. The outcrops of massive, compact dolostone on the open ridge north of road are mapped as unit c. Go north into woods, following broad ridge crest. Go west to the top of the next prominent ridge having steep cliffy eastern face. On top of the ridge (el. 770 feet on the topographic map) outcrop of cross-bedded grey dolostone and interbedded quartzite, showing that top is to the west, dips moderately west. The ledge on the east side of the ridge displays lithic characteristics of unit c very well. From here go east across gullies and up the main 840-ft. hill; rare outcrops are folded units c and d that on the whole dip west. Continue uphill; near the top of the hill are low and scattered outcrops of flat-lying unit e. Go southeast from top of hill to abandoned marble quarry on the southeast spur. In the quarry, interbedded calcareous sandstone and marble overlie massive coarse white marble; the calcareous sandstone sequence is best seen along the west rim of the quarry and is assigned to unit d; the main marble is assigned to unit e. Muscovite, phlogopite, and tourmaline are common in the marble. Notice the large recumbent fold and associated minor recumbent folds on the north face of the quarry; the folds have a movement sense consistent either with the normal limb of a fold and west-to-east movement sense, or with the inverted limb of a fold and east-to-west movement sense. Because the regional movement sense is east-to-west, the folds support the interpretation, based on lithic succession, that the sequence here is inverted. The recumbent folds are thus a record of the early deformation, which agrees with the fact that the early deformation in the trip area as a whole shows a recumbent style of folding. Between the outcrop of crossbedded dolostone and quartzite and this quarry, the axial surface of an early fold must exist, but its location cannot be precisely fixed.

The quarry is located in the core of a late anticline; notice that the beds here begin to assume an easterly dip. Farther to the east, the outcrops across the swamp in the woods are unit c again, dipping generally gently east and culminating in a late syncline containing units d and e. In between, however, there exists a zone of steep to vertical bedding dips; this is interpreted as the hinge zone of the early recumbent syncline (topping fold).

Proceed southwest from the quarry. Silty marble and dolostone of unit d showing nearly east trend and gentle south dip are exposed on the scuth slope of the 840-ft. hill. Continue southwest in wooded area and reenter the zone of west-dipping unit c just before the meadow. Return to car.

13.6 Proceed to driveway about 1,000 feet ahead to turn around. Return to Route 41; same corner as mileage 13.2. Turn right to go north on Route 41.

Mileage

13.7	Berkshire School; lunch on soccor field. Please--no litter! Carry all trash back to car. We are allowed to use the field for lunch through the kindness of the school authority; please do not abuse this privilege. Proceed north on Route 41.
15.0	Outcrop left of the road is unit e of the Stockbridge Formation.
15.5	Outcrops west of the road are unit e. East of the road, unit f and unit e crop out, the relations are interpreted as a thrust fault.
15.8	Large outcrops in pasture to the east (right) are unit f.
16.9	Junction of Route 23 and 41, same as mileage 7.3. Road cuts are nearly flat-lying calcareous sandstone beds of unit f, here in a series of imbricate thrusts. The same beds crop out in the stream to the southeast, and continue on to become the beds of unit f visited at stop A2.
17.6	Proceed straight north on Creamery Road where Route 23 bears to the right.
18.2	Sharp turn right onto McGee Road.
19.1	Intersection with Route 71; continue straight ahead on road bearing sign "Pumpkin Hollow Road."
19.3	Bridge over Green River.
19.4	Junction with West Plains Road. Continue straight ahead on Locust Hill Road.
19.6	Road cuts are in unit b of the Stockbridge Formation.
20.2	Road corner; big outcrops in garden to the left are in unit b3.
20.5	Road junction; turn left. Big outcrops are in unit b1; outcrops on the hill to the east across the pasture are in unit a.
20.7-21.0	Outcrops on the hill to the left are in units b1, b2, and b3, intricately interfolded.
21.4	Seekonk Road intersection. Turn right.
21.5	Red barn north of road. Stop A5.

Units a and b of the Stockbridge Formation. Park car near barn. Start traverse just behind (north) of the barn. The rocks are interbedded cream, pink, grey, and yellow dolostone, silty dolostone, grey silty phyllite, and arkose that weathers to a green cast; this is typical of the highest submember, b3. The beds dip to the east; they have been folded at least twice. The next rock type is a silty grey massive dolostone without interbedded arkose or phyllite. This is the middle submember, b2. The contact appears gradual. Continue east. This submember is in turn succeeded by a more heterogeneous unit, the lowest submember, bl, very similar to b3. These three submembers can be mapped over large tracts of the trip area and where distinguishable help to define the structural pattern.

Unit b is terminated to the east, along this traverse, by a feldspathic black phyllite with interbedded blue-grey silty limestone. The rock is identical with lower parts of the Walloomsac Formation, and is so mapped, even though the limestone has not yielded fossils, because this rock type is not found in the Stockbridge stratigraphic sequence. The area of black phyllite is limited. To the east, white, massive dolostone, unit a of the Stockbridge, is exposed, here much granulated apparently because the contact against the Walloomsac is a fault.

The contact between units a and b south of this area is gradational by addition of silty layers to the upper beds of unit a. The sequence is in the inverted limb of a major early anticline (bottoming fold). The Walloomsac is here interpreted to be in a graben. It is a relict either of an unconformably overlying sequence or of a thrust slice. Similar remnants of the Walloomsac are found commonly in the trip area. If the first interpretaticn is correct, then the overturned fold in the Stockbridge must be pre-Walloomsac. The writer, however, favors the idea that the Walloomsac, before the development of the graben, was in thrust contact with the Stockbridge. This interpretation has precedents; thrusting of flysch sediments over older carbonate rocks now characterizes much of the subalpine and prealpine zones of the Swiss Alps.

Turn around in barnyard and proceed west on Seekonk Road.

21.9 Road junction; bear right.

22.5 Road junction; bear right.

22.7 Road junction; continue straight ahead.

22.8 Road cuts are calcareous sandstone beds of unit d of the Stockbridge Formation.

Mileage

24.3 Black phyllite crops out in the road; this is the Walloomsac Formation. In the woods to the north are outcrops of carbonate rocks intricately folded with the allochthonous Everett Formation; the carbonate rocks are interpreted as constituting a sliver at the base of the allochthon. Hill south of road is likewise underlain by the Everett Formation.

24.7 Unit e of the Stockbridge Formation.

24.8 Allochthonous Everett Formation. Entering State of New York.

25.4 Intersection with Route 71; turn right.

25.5 Turn left onto Overlook Drive in front of the Green River Hotel and Bar.

25.7 Bridge over Green River.

26.7 Road junction. Continue straight ahead.

27.1 Microwave transmission tower. Stop A6.

Tectonic contact between the Everett Formation and the Walloomsac Formation. Carbonate slivers. Immediately west of the structure housing transmission facilities of the TV tower, green muscovite-chlorite-plagioclase phyllite is exposed; this is part of the main mass of Everett Formation. In the road, black phyllite characteristic of the Walloomsac Formation can be seen. East of the road, low rock outcrops at the north end of large pasture are a jumble of dolostone, limestone, and black and green phyllites; the carbonate rock types can be identified with units of the Stockbridge. Black and green phyllites occur between this outcrop and the next large outcrop of grey marble at the 1,370-foot elevation on the northern ridge of the main hill. The limestone is severely deformed resulting in a laminated S_2 fabric; microscopically the layers are alternatively relatively coarse calcite and extremely comminuted calcite. Note the strike of S_2 conforms to the contact with the main body of the Everett Formation (quartz-chlorite-muscovite-albite-pyrite), exposed immediately to the southeast. Go south along the steep west face of the hill. Warning: It's easy to get a sprained ankle unless you watch your step. Along this scarp, outcrops are nearly continuous for about 1,000 feet (300 m). The following are among features to be observed: The carbonate occurs in a zone varying rapidly from 0 to 10 feet (3 m) thick. Several rock types occur in the carbonate zone, locally obviously as blocks a fraction of an inch to several feet across. Presence of Everett Formation as blocks in the carbonate. Peculiar knobbly weathering surface of the dolostone, due to shattering of the rock later healed by quartz-calcite vein (fig. 3). Large solution cavities in the carbonate, part of which at least are natural and caused apparently by concentration of solution activity along boulder boundary. Repeated interlamination of black-green phyllite, interpreted to be caused by shear movement along Walloomsac-Everett contact. Recumbent fold of layers of limestone (best seen at the south end of the outcrops) whose axial surface passes into the overlying green phyllite to become the early foliation, showing that the foliation is no older than the emplacement of the allochthon and the carbonate sliver. Intense shearing of the phyllite (microscopically, the shear lamination is preserved as dusty inclusions in mica and plagioclase alike; these inclusions are in proper mutual orientation and at least the plagioclase microporphyroblasts grew after the development of the lamination).

The carbonate slivers are found in many places along the Walloomsac-Everett contact; the phyllite near the contact is always intensely sheared (the carbonate does not crop out well; more than once the characteristic intense shearing of phyllite has guided successful searches for the carbonate). The carbonate is assigned to various units of the Stockbridge, and their presence, the shearing of the adjacent phyllite, and the repetition of black and green phyllite along such contacts, have been used as evidence for assigning the Everett to an allochthon, as discussed in the main text.

In the pasture below (to the west of) the scarp, a second, discontinuous zone of carbonate occurs; the northernmost outcrop was seen at the beginning of the stop. From the scarp to the road the area is underlain by green and black phyllite on such a scale that they cannot be separately mapped; the entire zone is one of tectonic breccia. West of the road, the Walloomsac crops out; in turn it is thrust over green and purple slate, here assigned to the Everett Formation, but it may belong to the structurally lower Chatham slice (Zen, 1967). On the hill east of the carbonate zone, the green Everett continues and is part of the main outcrop belt of the Everett here.

Return to the car and proceed in the same direction. Steep down hill.

28.2 Junction with paved road. Bear right.

29.0 Junction with Route 22. Turn right, going north on Route 22.

29.3 Outcrops to the left belong to the Walloomsac Formation.

29.5 North Hillsdale.

29.6 Stop A7.

Unconformity of Walloomsac on the Stockbridge Formation; fossils in the Walloomsac; lithic types in the Walloomsac. Go east across field just north of apple orchard toward wooded hill (elevation 870 feet). At southeast spur of hill, quarry in massive grey dolostone, locally bearing black chert. This is unit c of the Stockbridge. At southeast wall of quarry, the dolostone is succeeded upward by 3-foot (1 meter) layer of dolomite-mottled grey marble, followed by black, silty thin-bedded limestone mapped as basal limestone of the Walloomsac. Go southeast about 400 feet (130 meters) to brook; in brook, black phyllite (Walloomsac) is exposed. On east bank, massive grey dolostone of unit c and calcareous sandstone of unit d rest on the Walloomsac on a thrust contact. Farther up hill unit e is next exposed.

Follow brook north. About 600 feet (200 meters) on, due east of 870-foot hill, black silty limestone like that in the quarry is exposed in the brook. Unit c is exposed locally in the brook as an inlier below the unconformity. Farther north in the brook the black limestone is fossiliferous. Apart from pelmatozoan fragments, a hash of gastropod and brachiopod fragments is preserved. There are no outcrops immediately east of the brook; the next outcrop on the slope is unit d of the Stockbridge again.

The relations at this stop are thus basically simple: Walloomsac, with basal fossiliferous limestone, unconformably on unit c. A thrust fault having an undetermined but probably minor stratigraphic throw locally brings unit c over the Walloomsac. The stop demonstrates the reality of the pre-Walloomsac unconformity, the stratigraphic near-identity of the Stockbridge on the two sides of the Taconic Range (unfortunately time does not permit us to see the other rock types of the Stockbridge west of the Taconic Range, but all units, except the Cheshire, Dalton, and possibly unit a of the Stockbridge have been identified in the proper sequence), and the age of the Walloomsac.

Return to the car. End of trip 3A. Those who wish to go directly to Albany could follow Route 22 north to the Berkshire Spur of the New York State Thruway in Austerlitz (exit B3). Go west across the Hudson River; take exit 24 to headquarters motel. Alternatively, one could pick up Route 203 in Austerlitz, go west to Valatie, and take Route 9 north from there to Albany. Those who wish to return to Great Barrington could either turn around, follow Route 22 south to Route 23 in Hillsdale, and go east on Route 23 to Great Barrington, or continue north on Route 22 for about 3 miles (5 km) to junction with Route 71, follow Route 71 south to Great Barrington.

ROAD LOG, TRIP 3B

Mileage

0 Start of trip at parking area just south of Route 23, 0.1 mile east of the New York-Massachusetts line and 2.9 miles east of junction of Route 22 and 23 in Hillsdale, New York.

Stop B1.

Walloomsac Formation. At least two sets of cleavage are evident in the roadcut; the late cleavage, S_3, folds the early cleavage, S_2 to produce an apparent east-over-west movement sense, most readily seen at the east end of the cut. Average attitude of S_2 is 25° azimuth and 55° dip SE, and the vein quartz is parallel to it. Average attitude of S_3 is 10° azimuth and 70° dip SE. The intersection of S_2 and S_3 produces a gently south-plunging crinkle lineation. There is, however, yet another down-dip crinkle visible on S_3; its origin is unclear. Mineral assemblage of the Walloomsac Formation at this outcrop is quartz-plagioclase-muscovite-chlorite; no carbonate has been found. Although the outcrop is located within the belt of retrogression discussed in the text, the rock does not show the effect because the carbonaceous phyllite of the Walloomsac Formation does not favour the development of even microscopic porphyroblasts.

Proceed east on Route 23.

0.1 Top of the Taconic Range; road cuts are all in the Walloomsac Formation.

0.7 Green phyllite of the Everett Formation at the south end of a narrow late syncline; from here south the syncline expands into the main area of the allochthon. Notice that one outcrop of the green phyllite shows west-dipping early foliation; this is interpreted as the result of refolding during the formation of the late syncline.

0.8 More black phyllite of the Walloomsac Formation and green phyllite of the Everett Formation.

0.9 Carbonate rocks belong to the Stockbridge Formation.

1.0 Turn right onto Jug End Road.

1.5 The ledges in the woods to the right (west) of the road are the Everett and Walloomsac Formations. The Everett is interpreted as debris left behind during the overthrust. The Walloomsac could be a thrust sliver; alternatively, the contact of the Walloomsac against the Stockbridge could be the Middle Ordovician unconformity.

Mileage

1.8 Left of road, behind barn, are calcareous sandstone beds of unit f of the Stockbridge Formation.

2.1 Intersection with Mount Washington Road. Turn right.

2.4-2.5 Outcrops along the road are unit g of the Stockbridge Formation.

2.6 Beginning of cuts in green phyllite of the allochthonous Everett Formation.

2.8 Stop B2.

Everett Formation and Egremont Phyllite in the zone of retrograde metamorphism. The large roadcut includes a variety of rock types that characterize the Everett Formation in areas of medium-low metamorphic grades: Micaceous phyllite, feldspathic phyllite, quartzose phyllite. These rocks are predominantly green, but purplish green varieties can be found, e.g. near the west end of the cut. The rocks show at least two sets of cleavage: S_2, a foliation, is at least locally parallel to bedding, S_1, and can be best seen near the middle one of three concrete storm-sewer covers situated on the north side of the road cut. Here S_1 and S_2 are nearly vertical but S_3 dips east; an apparent west-over-east movement sense results. At one place, the axial parts of S_2 folds can be seen to be refolded by S_3, and the early axes trend diagonally across the late limbs to produce a rhombic interference pattern.

The mineral assemblage here is: Quartz-chlorite-muscovite-plagioclase-magnetite. Magnetite is most readily seen near the west end of the cut, and is in both green and purple (hematitic) phyllites. The plagioclase is zoned, varying from about An_{15} to An_0 according to electron microprobe data, and shows retrogression as discussed in the main text. Notice that the magnetite is also retrograded to chlorite.

From the west end of road cut walk 500 feet (150 m) west to private drive and small bridge across Karner Brook. In the brook is exposed black phyllite typical of the Egremont Phyllite. Some of the phyllite layers are calcareous. Notice the presence of two prominent sets of cleavage again; folding of S_2 by S_3 led to the steeply south-plunging cleavage folds. For reasons discussed in the main text, the Egremont Phyllite is identified with the Wallomsac Formation; compare the outcrop with the outcrop of Walloomsac at stop B1. In fact, the Egremont Phyllite of stop B2 probably is geometrically continuous with the outcrop of Walloomsac at stop B1; these two areas of black phyllite are separated only by a ridge of green phyllite of the Everett Formation, and the green-black contacts occur at about the same elevation on the two sides of the ridge.

Three hundred feet (100 m) downstream, the Everett Formation crops out in the stream bed, thus in map pattern the Egremont Phyllite is completely surrounded by the Everett Formation. Unfortunately, the contact is not exposed here. Because as a whole the Everett occupies high ground and the Egremont occupies lowland, and because in rare exposures of the contact the Everett overlies the Egremont, the Everett is interpreted as geometrically above the Egremont, and the fact that the Everett here crops out 300 feet downstream from the Egremont is explained by local irregularities of the surface of the contact. This contact is considered to mark the base of the allochthon; however, unlike the Everett-Walloomsac contact in much of the Egremont quadrangle (e.g., stop A6), the Everett-Egremont contact is so far not known to be marked by carbonate slivers.

Proceed west on Mount Washington Road.

3.0-3.9 Green and minor purple beds of the Everett Formation; mostly muscovite-albite-chlorite-quartz phyllite containing hematite or magnetite. The Egremont Phyllite occurs almost exclusively in the deep ravine of Karner Brook east (left) of the road.

4.9 Road junction; bear left with paved road.

5.0 Gully crossing. Contact of Everett Formation and Egremont Phyllite has climbed up to this point. The flat area ahead is underlain by the Egremont Phyllite whereas the higher hills surrounding the plain are underlain by the Everett Formation.

5.7 Road corner; turn left with paved road.

7.8 Junction of road to Mount Everett, the highest peak in the southern Taconic Range. On a clear day a fine view can be had from the top.

8.1 Road intersection; continue straight ahead. This is downtown Mount Washington.

9.3 Road intersection, stop B3.

Medium-to-low-grade Everett Formation. Rocks immediately east of road in the woods, opposite gate on Mt. Plantain Road, are chloritoid-muscovite-chlorite-ilmenite-plagioclase schist of the Everett. We are just north of the almandine isograd here. The rock has been separated and analyzed; K/K + Na ratio in the muscovite = 0.8 and d_{002} = 9.95; Ab/Ab+An in plagioclase is about 0.8, but the plagioclase is weakly zoned (no retrograde zoning). The ilmenite is nearly stoichiometric $FeTiO_3$ with minor MnO. In the chlorite, Fe"/Fe"+Mg+Mn+Ca ratio is 0.67

and Mg/Fe"+Mg+Mn+Ca is 0.30; in the chloritoid, Fe"/Fe"+Mg+Mn = 0.84 and Mg/Mg+Fe"+Mn = 0.12

Notice the outcrop of the black Egremon Phyllite in the road.

Proceed south on main gravel road.

9.6 Outcrops below red barn to the west are calcareous phyllite beds of the Egremont Phyllite.

10.0 Back into the Everett Formation; the Egremont Phyllite occupies the lowland on the left (east). Except for one more brief skirmish with the Egremont Phyllite, we will stay in the Everett Formation for the next several miles. We are near the southern terminus of the inlier of Egremont Phyllite.

10.5 Here in the road is the aforementioned brief skirmish with the Egremont Phyllite.

10.8 Mount Ashley to the right; paragonite-bearing chloritoid schist of the Everett Formation.

12.0 Entering Connecticut.

14.8 Stop B4 at parking area. Follow blue blaze to top of Bald Peak.

Medium-grade Everett Formation. View of the Taconic upland, including Bear Mountain to the north, the highest peak in Connecticut, and Mount Frissell to the northwest, whose south slope includes the highest point in Connecticut. Lions Head, the sharp ridge to the ESE, is our next stop. The topographic contrast between the allochthon and the autochthon (lowland) is obvious. Notice in the outcrop the presence of three S surfaces in addition to the bedding: S_2, the early foliation, apparently conforms to the bedding in flat outcrops; however, the vertical outcrop faces along the trail near the top show recumbent folding of bedding having S_2 as the axial surface. S_3, the dominant cleavage, folds S_1 and S_2 to produce a characteristic south-plunging set of folds. S_4 is the latest cleavage; it is nonpenetrative and produces sharp chevron folds of S_3. The rock here is an almandine-chlorite-muscovite-plagioclase schist; the plagioclase and quartz locally are granulated and the plagioclase not obviously zoned. Chloritoid is confined to inclusions in the almandine; staurolite has not been found here. Some magnetite-rich layers are without almandine.

Bald Peak has a tangled nomenclatorial history vis-a-vis its geology. Hobbs (1893a, p. 3) called it Mount Riga, and designated it as the type locality of his Riga Schist, which from his map pattern must be identified largely with the Walloomsac. The hill has since been renamed, and the rock is now mapped as the Everett. However, another peak farther south, in the Sharon quadrangle and visible on the skyline is now called Mount Riga on the U.S.G.S. topographic sheet, and it is indeed underlain by the Walloomsac!

Return to car and continue south.

15.3 Road corner; bear left with main road. Pond just before the corner is South Pond, a private preserve. Stone structure to the right just beyond the corner is a reconstruction of a blast furnace operated by one Joseph Pettee during the early years of the 19th century. The iron ore was hauled up from the valley; many of the pits are still prominent landscape features near Salisbury, Conn. (The ore was developed along the contact of the Stockbridge and Walloomsac Formations, presumably a record of pre-Walloomsac subaerial erosion.) Charcoal for the furnace was locally obtained in the mountains; many platforms where charcoal was produced can still be recognized in the woods today.

15.6 Crossing Wachocastinook Brook which drains South Pond.

16.1 Waterfalls in the brook are visible; almandine-staurolite-plagioclase-chlorite-muscovite-quartz-ilmenite-armoured chloritoid schist of the Everett Formation.

16.4 Lions Head is the sharp ridge to the left. Along the road are cuts in the Everett Formation. The contact with the underlying autochthonous Walloomsac Formation is just ahead, but is not exposed.

17.4 Bridge over the Wachocastinook Brook. Staurolite-almandine-biotite-muscovite-plagioclase-quartz-ilmenite schist of the Walloomsac Formation in the stream.

18.2 Road comes in from the right. 200 feet ahead, Bunker Hill Road comes in from the left. Turn sharply left onto Bunker Hill Road, going north. We are now in unit e of the Stockbridge Formation.

18.5 Road fork; bear left with the main road.

19.2 End of road; stop B5.

Medium-high-grade Everett Formation and Walloomsac Formation. From the house follow the white-blazed Appalachian Trail northwest for a 0.7-mile (1.2 km) walk to Lions Head (total climb about 550 feet, 180 m). View of the autochthonous lowland; the hills along the skyline to the east are in the Berkshire massif. Note the steep scarp of Canaan Mountain to the southeast; this is the type locality of Canaan Mountain Schist of Rodgers and others (1956), a controversial sillimanite-muscovite schist of uncertain age which the writer believes to be mostly high-grade Walloomsac Formation.

On top of the lookout point of Lions Head, coarse-grained staurolite-almandine-chlorite-plagioclase-muscovite schist of the Everett Formation. Fe/Fe+Mg in the chlorite is about 0.6; in the almandine, about 0.86; in the staurolite, about 0.83. A notable feature is enrichment of Zn in the staurolite--about 0.3 wt. percent of ZnO was reported in the analysis of mineral separate. Electron-probe analysis has not shown any zoning of the staurolite; in the almandine, Mn shows mild enrichment in the center, whereas Fe shows a corresponding enrichment in the rim, as might be expected; Ca shows no zonal distribution. Agar (1932) reported kyanite in the general vicinity, but it has not been found in the present study. The grade of metamorphism is consistent with steady increase from chlorite phyllite in the Bashbish Falls area to sillimanite-biotite-almandine-muscovite schist and muscovite-microcline-almandine-biotite schist in the Canaan Mountain region.

Note the presence here of early recumbent folds in S_1 having S_2, the early foliation, as the axial surface. S_2 is folded by S_3 to produce the prominent south-plunging folds.

Return to car via Appalachian Trail. Time permitting, we will go east to low knob in the woods, east of the house and across the narrow part of the pasture. Outcrop just east of top (elevation 1,200 feet) is an almandine-staurolite-biotite-plagioclase-muscovite schist of the Walloomsac Formation; the almandine and staurolite, interestingly, have compositions indistinguishable from those of the Everett Formation at Lions Head. The Fe/Fe+Mg ratio in the biotite is 0.50. The K/K+Na ratio of the muscovite is 0.74, and d_{002} = 9.93. The muscovite and biotite from this outcrop have yielded the isotopic ages given in the main text.

20.1 Return to corner at mileage 18.2. Continue straight ahead.

20.6 Road cuts are rusty weathering staurolite-almandine schist of the Walloomsac Formation.

20.9 Junction with Routes 41 and 44. Village of Salisbury. Turn left onto Route 44, going north and east.

21.0 Go straight ahead on Route 44.

22.2 Road cut of Walloomsac Formation.

22.8 Nearly flat-lying calcareous sandstone and interbedded marble; unit d of the Stockbridge Formation.

23.1 Salisbury School.

23.4 Turn right onto dirt road ("Wildcat Hollow Road"). Outcrops on the main highway ahead are coarse staurolite-almandine-biotite-muscovite-plagioclase-ilmenite-quartz schist of the Walloomsac Formation.

Mileage

23.5 Walloomsac Formation. Relative to the Walloomsac of the next stop, these outcrops have been downfaulted.

23.9 Pasture east (left) of road with tracks leading up to it. Drive into pasture and park. Stop B6.

Unconformity of Walloomsac Formation on the Stockbridge Formation. Walk up the pasture east of the road. In the pasture, the low ledge is a buff-cream-weathering cream dolostone interbedded with silty phyllite, characteristic of unit b of the Stockbridge. Continue east into woods. Scattered outcrops are also unit b. Walk up to the base of the steep slope (elevation 950 feet); the outcrop is a massive grey dolostone typical of unit c of the Stockbridge. The contact of the two units shows a complex pattern. Measured axial-plane attitudes of minor folds and exposed contacts of units b and c all show that the axial plane of the larger fold dips east at low angle. Walk up the steep slope to the bluff at 980 feet. Here, grey massive dolostone is directly overlain by black calcite-biotite-plagioclase schist of the Walloomsac Formation. Erosion has exposed the contact in three dimensions; traces of bedding in the dolostone show up the angular unconformity. The origin of the silty material defining the bedding is uncertain: Could it be Walloomsac sediments deposited in weathered-out bedding cracks of unit c?

The unconformity surface has been folded into isoclinal folds whose attitude and style suggest the early, recumbent folds in both the Stockbridge Formation (e.g., stop A4) and in the Everett Formation (e.g., stops A6, B4, B5). The relations thus indicate that the early recumbent folding is later than the deposition of the Walloomsac Formation. (I am grateful to R. S. Stanley for first noticing these folds.)

At the north end of the bluff, fine mats of tremolite are developed as reaction rim along the contact. The top of the hill consists of the Walloomsac. The basal contact can be traced to the south and east, practically continuously for about 1/2 mile (1 km); at the east end of this contact the Walloomsac is directly on unit b of the Stockbridge.

Return to car, end of trip. Go north back to Route 44 for easy access to other major highways.

FIGURE CAPTIONS

Figure 1. Topographic map showing area of the Taconic Range in the trip area. The 7 1/2-minute quadrangles outlined are as follows (left to right): Top row, Hillsdale, Egremont, Great Barrington; middle row, Copake, Bashbish Falls, Ashley Falls; bottom row, Millerton, Sharon, South Canaan. Triangles designate stops for trip 3A; squares designate stops for trip 3B.............................. p. 3-35

Figure 2. Generalized geologic maps of the trip area. (a), the Egremont quadrangle and adjacent areas of the Hillsdale quadrangle from Zen and Ratcliffe (in press). (b), the Bashbish Falls quadrangle and adjacent areas of the Copake, Millerton, and Sharon quadrangles from Zen and Hartshorn (1966) and Zen (unpublished data). (c), a representative cross-section modified from Zen and Hartshorn (1966). Trip stops given by numbers 1A, etc.. p. 3-36-37

Figure 3. Polished sections of shattered-and-healed dolostone in carbonate slivers along the base of the Taconic allochthon of the trip area. The right-hand sample is from stop A6.................................. p. 3-38

Figure 4. Polished sections of stretched pebbles of the small-pebble conglomerate unit of the Everett Formation, from Bashbish Falls. Left, section parallel to elongation of pebbles; right, section normal to elongation of pebbles.. p. 3-39

Figure 5. Trip stops. (a), trip 3A stops, (b), trip 3B stops. From Zen and Hartshorn (1966); Zen and Ratcliffe (in press); Zen (unpublished notes). For letter symbols, see table 1. Scale is 1:24,000................. p. 3-40-41

0 5 10 Km

0 5 Mile

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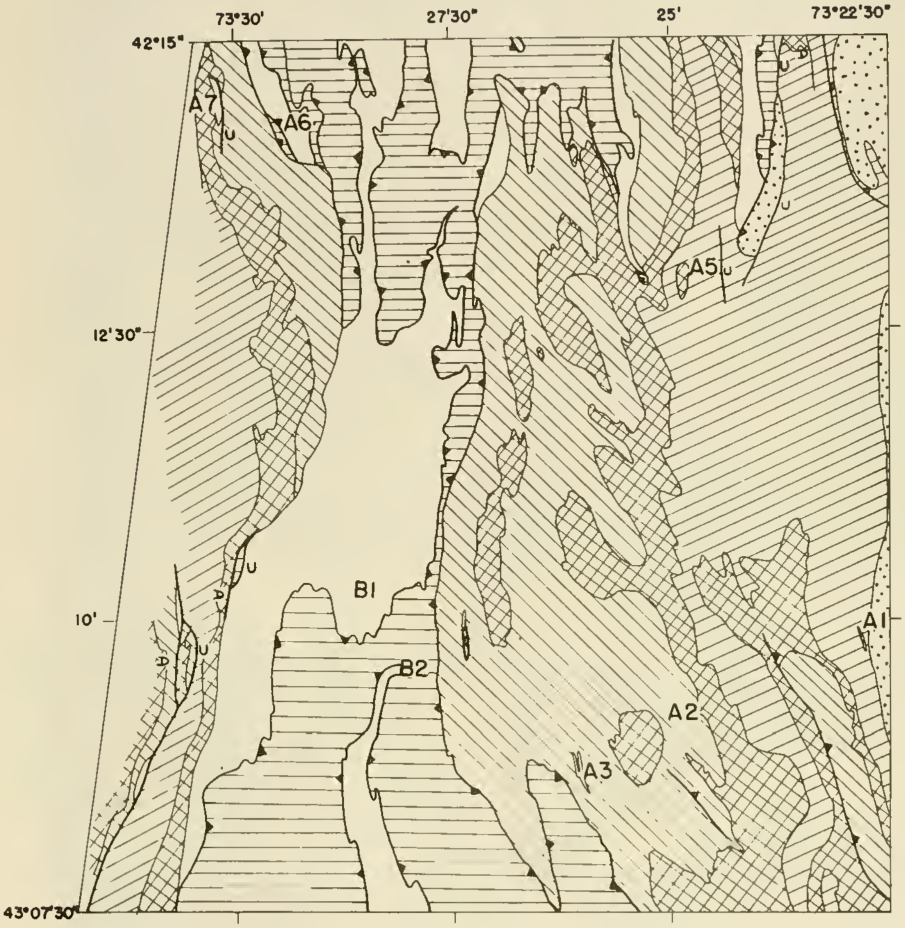
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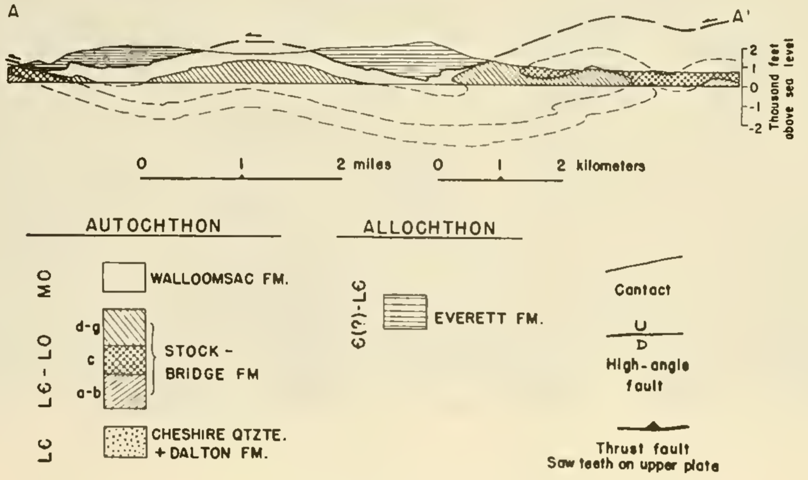
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STOCK - BRIDGE FM
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CHESHIRE QTZTE. + DALTON FM.
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Contact
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D
High-angle fault
Thrust fault
Saw teeth on upper plate

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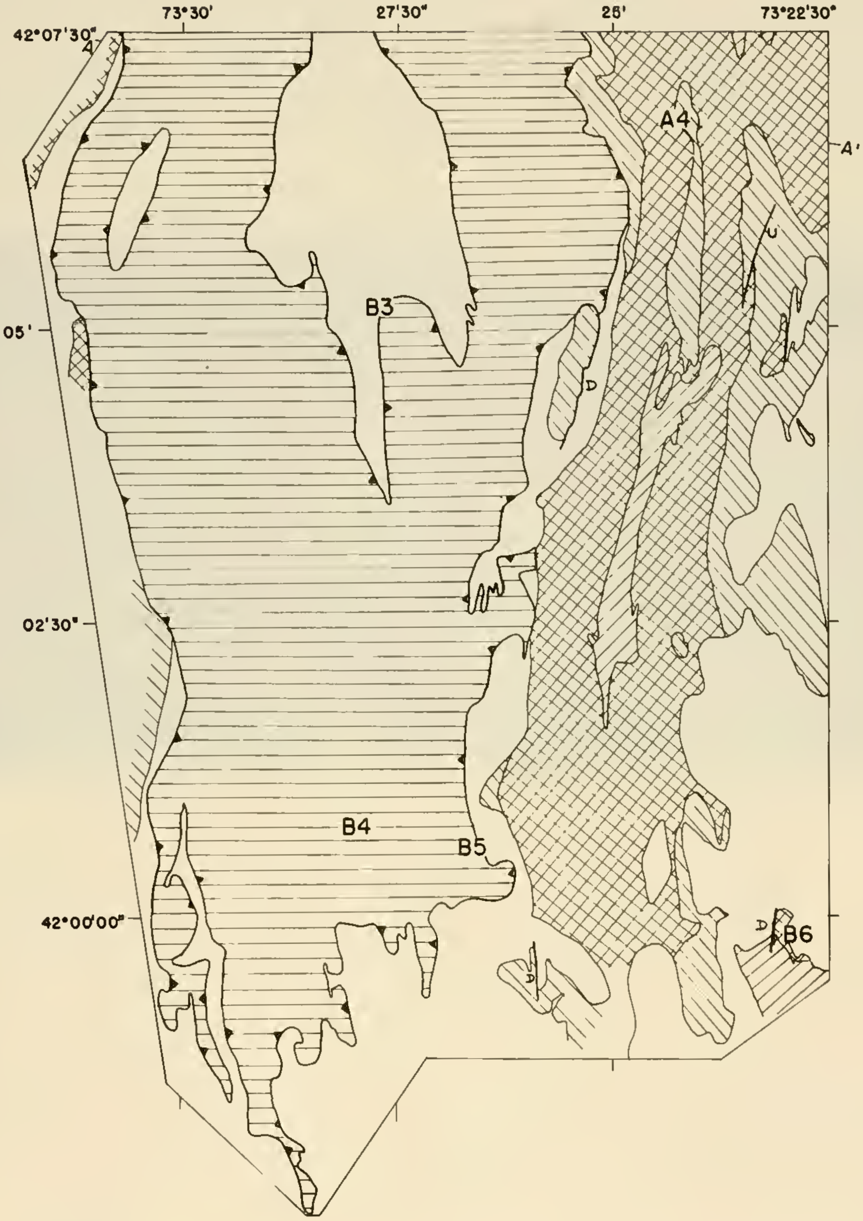
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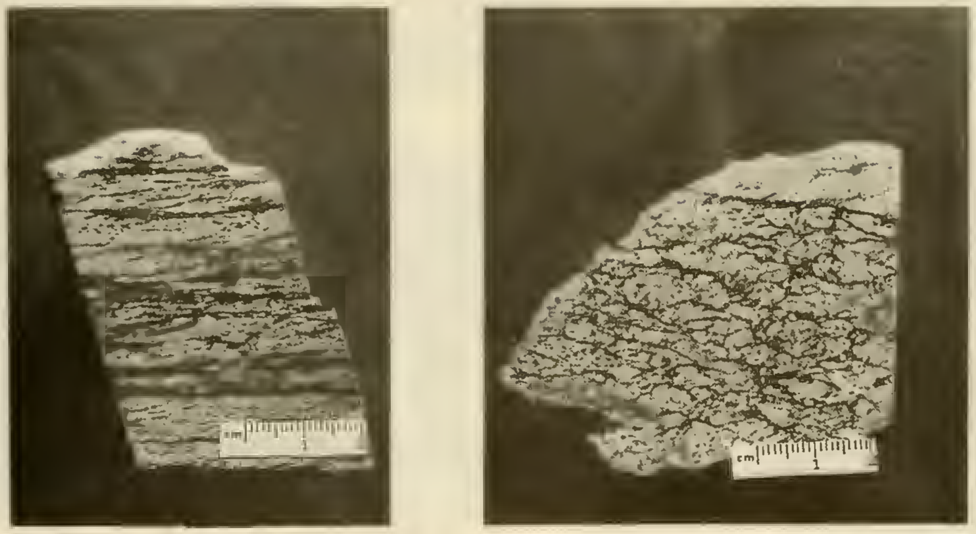

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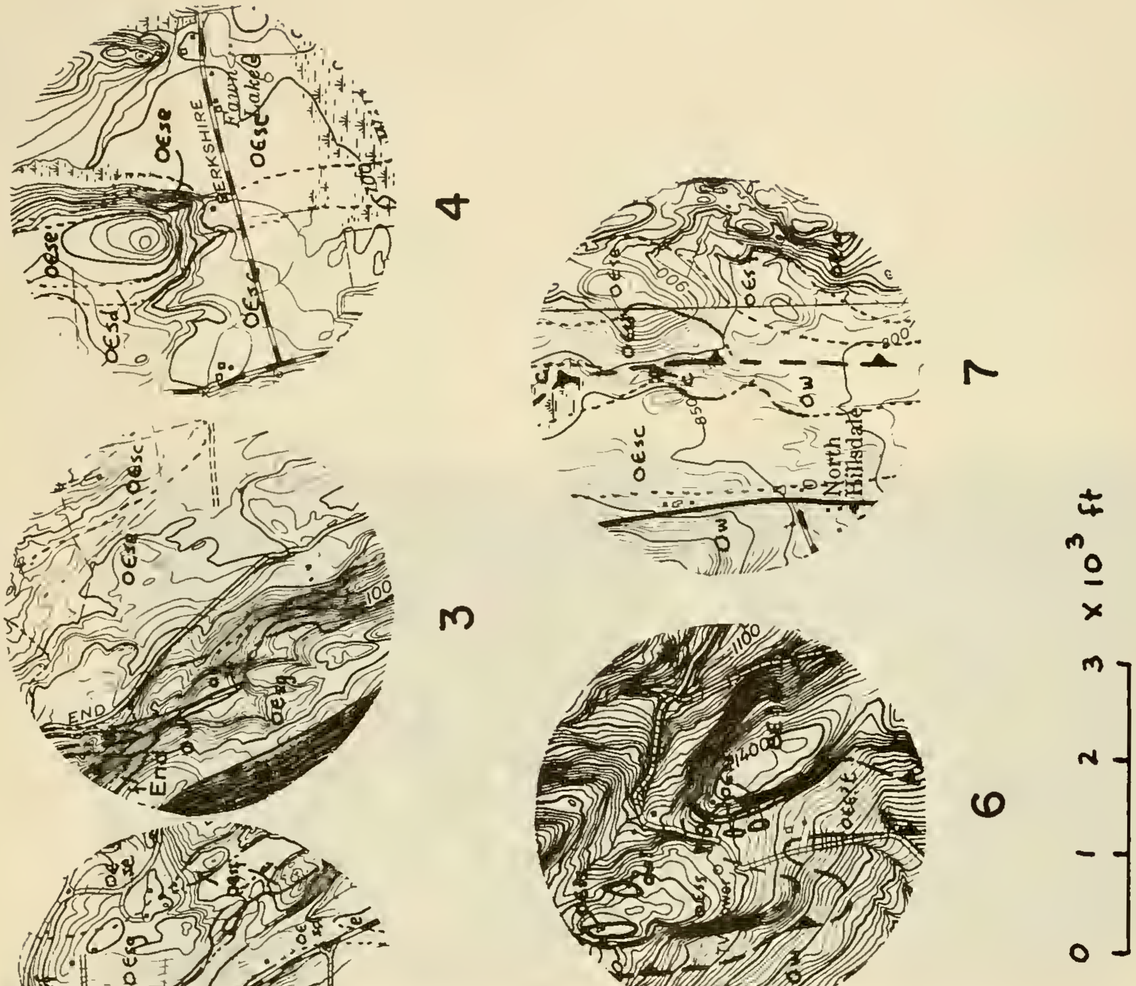

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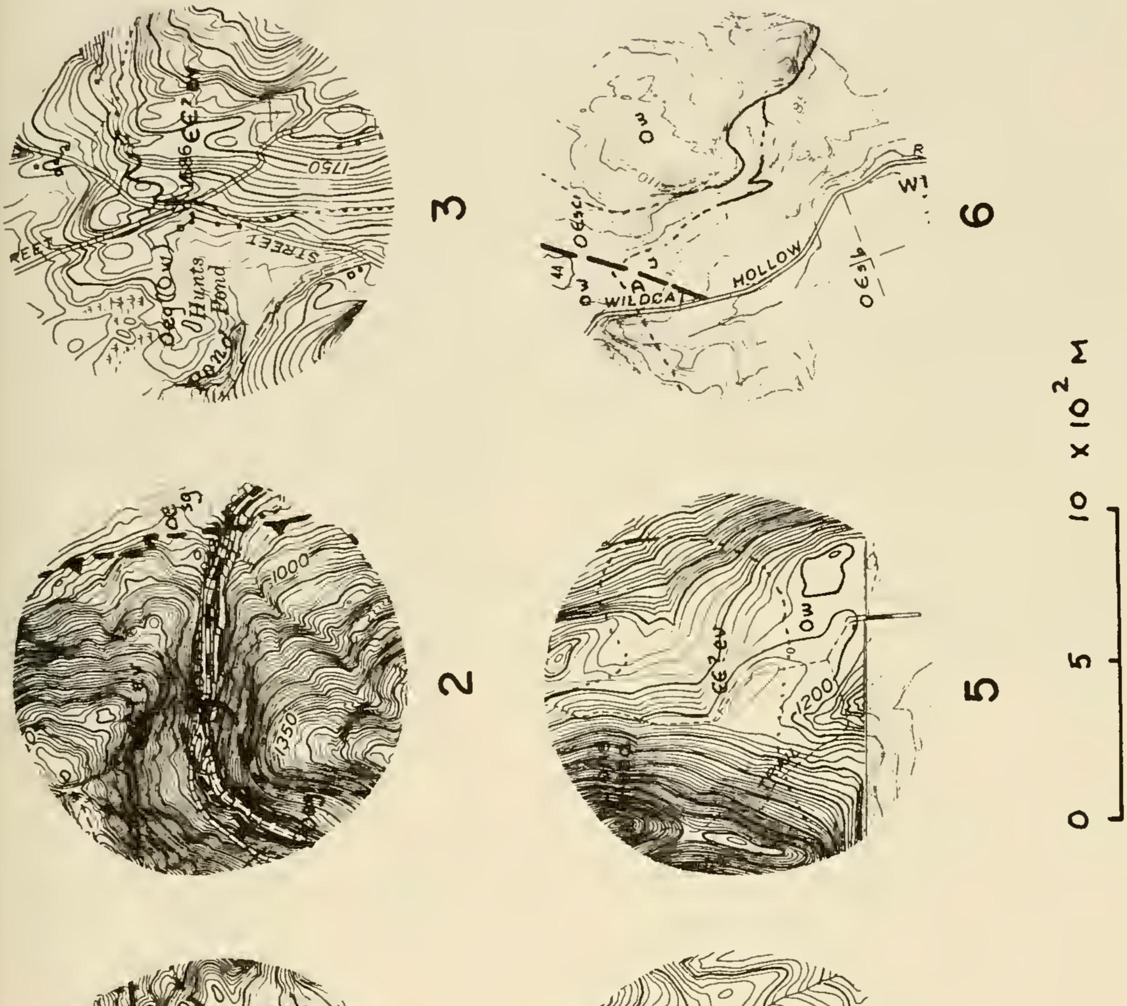
1
2
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Hynts Pond
STREET
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1800
1200
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1350
1750
900
0 5 10 x10² M

5b

TRIP 5

GLACIAL GEOLOGY OF THE SCHOHARIE VALLEY

by

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Rensselaer Polytechnic Institute

INTRODUCTION

Recent mapping, literature, and field trips through the Hudson-Champlain Lowland [LaFleur 1961, 1965, 1965b, 1965c, 1968, Schock 1963, Connally and Sirkin 1967, 1969, and Wagner 1969] have provided detailed analyses and interpretations of a variety of glacial deposits and have reevaluated the series of proglacial lakes which accompanied wasting Late Wisconsin ice. Readvances of ice in the southern Hudson and southern Champlain Valleys have been proposed by Connally (1968), and Connally and Sirkin (1969).

The Hudson Valley meltwaters seem to have always had exit to the south away from the receding ice, a situation which produced a rather consistent pattern of outwashing ice-contact, fluvial, and lacustrine deposition into Lakes Albany and Vermont, the levels of which were progressively lowered as crustal uplift proceded.

The Schoharie Valley was, however, a container of a slightly different character, draining northward toward the ice instead of away from it, having several sharply defined notches along its southern divide through which glacial drainage spilled, and also having along the northeastern divide several shallow cols through which reinvasions of Hudson Valley ice passed. These features made the Schoharie Basin extremely sensitive both to major ice advance and to more subtle action of wasting and readvancing ice, and permitted the accumulation in deep, sheltered valleys of a more complete Pleistocene record than has been previously seen in eastern New York.

Among the intriguing problems in understanding the deglacial history of the region are the determination of routes for exiting meltwaters issuing directly from wasting and readvancing ice, and confirmation of the presence of lakes standing behind spillways as postulated by earlier workers. Also more than 400 water well logs have been recently obtained from drillers who have worked in and near the Schoharie Basin, and many of these logs suggest earlier glaciations in eastern New York. The subsurface data have made it abundantly clear that the complete Pleistocene record will not be determined through surficial mapping alone. As is usually the case with drillers' logs this information is oversimplified, and variable in its detail and reliability, but it provides the only means of access to the older glacial deposits which rarely outcrop. The last major glaciation, of probable Cary age, was an overwhelming one, and subdued the record of any prior glacial episodes with a thick blanket of till, drumlins, and lacustrines.

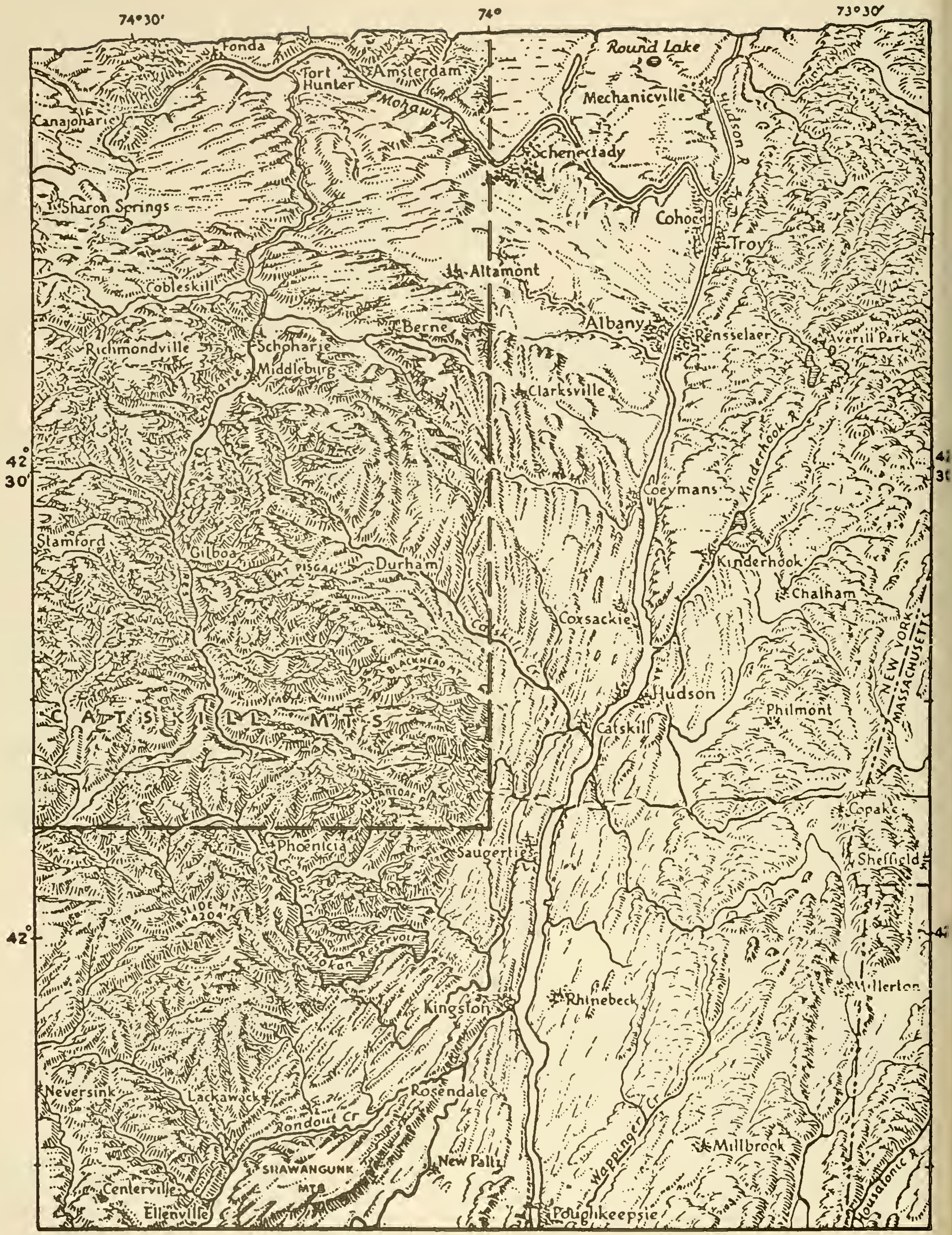

FIGURE 1. Physiographic diagram of Catskills and Hudson Valley showing area of Figure 3. (After Berkey, 1933)

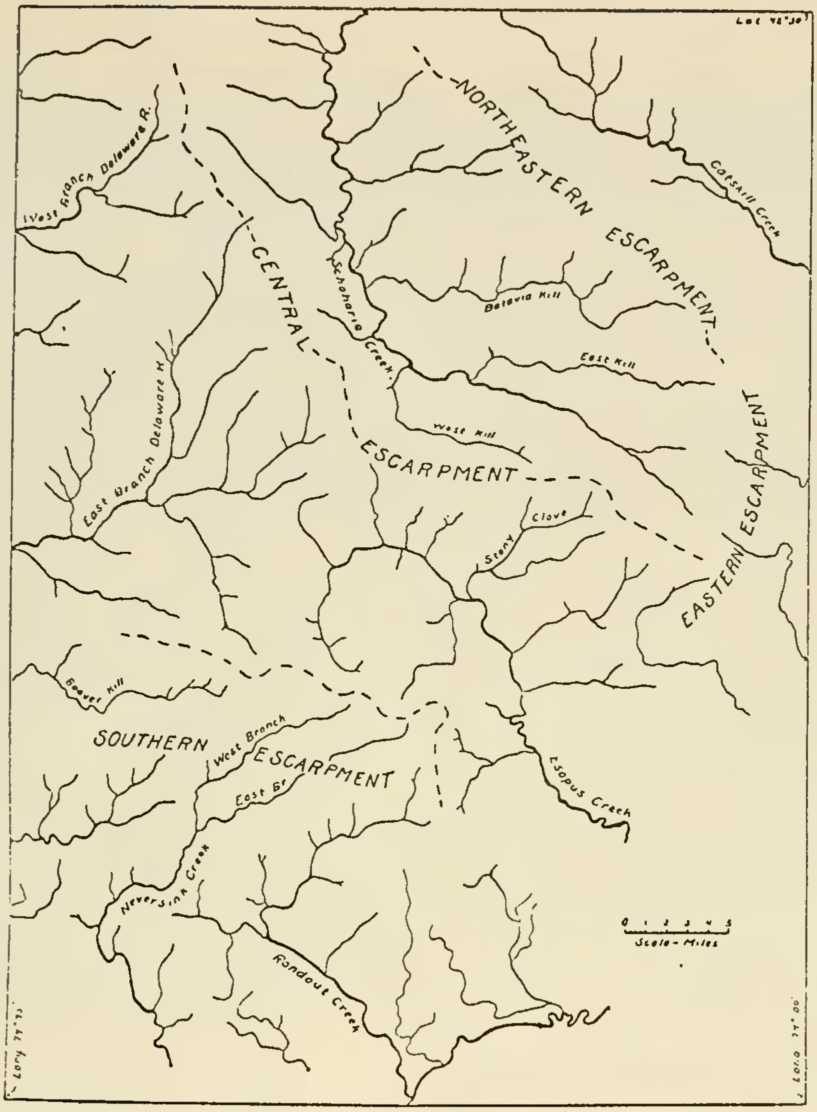

FIGURE 2. Drainage of Catskills, with escarpment terminology of Rich (1935)

SCHOHARIE BASIN PHYSIOGRAPHY

Schoharie Creek above Breakabeen and its tributaries drain an area about 15 miles wide and 40 miles long, situated between two northwestward-trending escarpments developed on gently dipping Devonian clastics. (See Fig. 2) The Central Escarpment extends from Kaaters Kill and Platte Kill Cloves northwestward through Hunter Mountain, West Kill Mountain, Vly Mountain, Irish Mountain, and the Moresville Range to a point between Stamford and Jefferson where it loses its prominence. This ridge forms the divide between the Schoharie and the Delaware-Susquehanna drainage. The North-eastern Escarpment forms the Schoharie-Catskill divide and extends from North Lake near Haines Falls northwestward through the Black-head Mountains, Mount Zoar, Mount Hayden, and High Knob toward Broome Center. A topographically reduced extension of the North-eastern Escarpment can be projected across the Schoharie Valley to

Summit, Richmondville, and South Valley. According to Rich (1935), the high peaks above 3000 feet in the Central Escarpment formed loci for dispersion of local mountain glaciation for a time during the disappearance of the continental glacier from the upper Schohar[ie] Basin.

The Schoharie-Hudson-Delaware divide passes along some of the highest mountains in the Catskills, broken by deep notches at only three points - Stony Clove (2070'), Deep Notch (1900'), and Grand Gorge (1570'). The western corner of the basin passes into the Susquehanna near West Richmondville through a notch and outlet channel at 1500'. Two other outlet channels along the eastern edge of the Schoharie Basin lead into the Catskill-Hudson drainage at Franklinton south of Middleburg (1180'), and into the Normanskill-Hudson drainage at Delanson (840'). During the glacial deglacial history all six of these outlets played important roles, but not all at the same time, and some more than once. Several cols may also be found along the Northeastern Escarpment, but none of these served as meltwater spillways; rather their function was to provide access for readvancing ice of the Hudson Lobe into the Schoharie Basin.

An important area of karst topography developed in the Helderberg limestones along their outcrop across the middle of the Basin between Sharon and West Berne. The influence of the cavern systems on deglacial drainage might prove very significant. The hydrology of the terrane is described by Baker (in preparation) who investigated several miles of underground chambers.

EARLY CONCEPTS - LAKE SCHOHARIE

Papers by Brigham (1908), (1929), Fairchild (1912), and Rich (1935), describe the glacial history of the upper and lower reaches of the Schoharie. Brigham (1929), and Rich mapped northward from the Catskills through most of the Gilboa quadrangle and Brigham included the lower Schoharie in part of his Fonda and Amsterdam sheets.

Much of the earlier Pleistocene literature concerned itself with areas where various glacial lake waters collected, and citations of cols which controlled their levels. The Schoharie Basin was no exception. Brigham (1908), first appreciated the presence of glacial and post-glacial lake water standing in the Schoharie Valley and gave these lake bodies that name. His post-glacial lake extended up the Cobles Kill and main Schoharie valleys and was dammed by a morainal plug at the northern end between Burtonsville and Esperance. The water level was set at 700 feet. Brigham[']s glacial Lake Schoharie was postulated as an ice-margin body draining through cols in the southern divide.

However, Fairchild's concept of Lake Schoharie was of a more expansive water body extending far outside the present basin. He included this lake in the middle of his system of falling water levels beginning with Lake Herkimer and ending with Lake Amsterdam. These sequential lakes presumably accumulated in the Mohawk Lowland and adjacent valleys between two opposed receding lobes of ice, the Ontarian lobe on the west and the Mohawk-Hudson lobe on the east.

Rich's Lake Schoharie was a two-phase water body which overflowed to the Delaware Basin first at the Grand Gorge col at 1620'-1570' and extended between these levels as far north as Middleburg. The Franklinton col, upon being uncovered by the receding ice controlled the final phase of discharge to the Catskill Basin at an elevation of 1180'.

LAKE SCHOHARIE DEPOSITS AND ATTENDING PROBLEMS

Large discontinuous masses of red and gray clay occur through the Schoharie Valley from Prattsville to Esperance, a distance of some 30 miles. The surface elevations of these clay deposits drop progressively from 1300 feet near Prattsville to a level of 650 feet at Esperance. Although considerable gullying has modified the original boundaries of these deposits there are many places where ice contact surfaces seem to be preserved. There are also several exposures which exhibit deformation of the clay by contact with the ice and also by ice override. The great bulk of the clay is confined to dissected terraces at the base of steep valley walls and beneath the valley floor alluvium. There are few places where high-level clays of any great thickness or extent occur. Even in these high-level cases it appears that the thin clays represent only the bottom deposits of small short-lived ice-border lakes.

Fairchild (1912), and Rich (1935), presumed that these clays were the bottom deposits of an expanding Grand Gorge Lake, which had exit through the divide into the Delaware Basin at a point two miles southwest of the village of Grand Gorge. The present base of the channel at the Grand Gorge col is 1570 feet. According to Rich, downcutting of the Grand Gorge outlet between the elevation of 1620' and 1570' took place while the Schoharie ice block was wasting northward to a point just south of Middleburg. When the receding ice supposedly uncovered the present col at the head of the Cats Kill drainage, at a point now between Middleburg and Franklinton, the Grand Gorge Lake waters abruptly fell, downcutting the Franklinton "till plug" at the divide to a level of 1180 feet. This explanation of the history of Grand Gorge Lake requires first that a reasonably continuous water body existed between Grand Gorge and Middleburg at a level between 1620 and 1570 feet. It also assumes that the valley-choking plug near Franklinton is older than the outlet at Grand Gorge. Serious difficulties arise with this hypothesis.


AMSTERDAM
SARATOGA CO
SCHENECTADY CO
MONTGOMERY COUNTY
SCHOHARIE COUNTY
SCHENECTADY COUNTY
ALBANY COUNTY
HELDERBERG
MOUNTAINS
MOHAWK
Schoharie
Creek
Esperance
Duanesburg
Delanson
Quaker Street
Mariaville
Berne
East Berne
West Berne
Knox
Middleburg
Cobleskill
Central Bridge
Carlisle
Sharon Springs
Sharon
Richmondville
Charleston
Burtonsville
Fonda
Glen
Canajoharie
Ames
Marshville
Seward
Hyndsville
Gallupville
Browns Hollow
Rural Grove
South Amsterdam
Cranesville
Thompsons Lake
Warner Lake
Summit Lake
Schuyler Lake
Bear Swamp
OTSEGO CO

FIGURE 3. Topography of the Schoharie Basin showing positions of the ice margin and locations of trip stops.

If an expanding lake controlled by a bedrock sill at Grand Gorge existed, one should anticipate abundant evidence for deltas in all of the major tributaries at elevations between 1620' and 1570'. Side tributaries between Prattsville and Middleburg are indeed partly filled with fluvial deposits but all of these show indications of having been laid in close proximity to ice rather than as deltas extending into an open lake. One would also expect to find more uniform summit elevations for the top of the clay surface, such as that of Lake Albany, through the length of a large lake basin rather than the present configuration of the clay surface which is step-like and in many places reflects contact with the ice. If Rich's hypothesis is true then the clays two miles southwest of Middleburg at the north end of Grand Gorge Lake must have been deposited in over 800 feet of water. Clays should have also extended in such a deep lake well up into the tributary valleys. Evidence for their presence in these valleys again is associated quite distinctly with ice blockage rather than as continuing masses connected to the clay deposits in the valley bottom. Finally the sudden onslaught of Grand Gorge Lake waters upon the valley plug at Franklinton would certainly have caused more erosional damage than the Franklinton plug seems to have experienced. Fairchild (1912), considered that the Franklinton channel carried Cats Kill drainage northward into Lake Schohar because of his impression of slight erosion there. The field evidence can be better applied to an alternative hypothesis.

MULTIPLE GLACIATION

Rich (1935) correctly recognized abundant field evidence for readvances of the wasting ice in the upper Schoharie Valley, but because his mapping terminated in the Gilboa quadrangle, he could not apply this concept to the complete deglacial history of Lake Schoharie. There are also many kame terraces and deltas which occur throughout much of the upper Schoharie Valley at elevations below the 1570' threshold of the Grand Gorge outlet. At Prattsville thick lacustrine clay overlies coarse gravel. These situations seem to require either draining of the Schoharie Valley and later impounding of lake waters by readvances of the ice, or else very leaky ice through which northward drainage was relatively easy. Evidence to date strongly favors the readvance hypothesis. The Franklinton plug seems to have been emplaced in part by one of these late readvances. Therefore its age is critical to the history of Lake Schoharie through the Grand Gorge phase. If the upper Cats Kill were free of ice and sediment blockage it could easily have captured the waters of Grand Gorge Lake as the ice receded, lowering the level to below 1100'. A readvance to the vicinity of Franklinton and to Breakabeen would impond Schoharie waters again abut at a lower elevation. This process may have operated several times prior to the final construction (and destruction) of the Franklinton plug, for it appears to have a compound character. It is certainly not all till as Rich assumed.

Four positions of the ice margin are recognized as significant in the deglacial history of the Schoharie Valley. See Figure 3. It is probable that all four represent terminal positions of the ice during readvances which occurred during Late Cary(?) deglaciation. These readvances are particularly important to the lake history which heretofore has been treated as a simple recessional sequence.

Tannersville Margin

The oldest readvance is represented by the ice margin which extends along the edge of the Central Escarpment. The leading edge of this advancing ice lobe moved eastward in the Hunter-Tannersville area where it apparently collided with a lobe of Hudson ice advancing up the Kaaters Kill Clove at Haines Falls. In several water well logs between Hunter and Tannersville an upper till some 15 to 30 feet thick can be recognized overlying a much thicker section of lacustrine sediments. A basal till lies directly beneath the lacustrines on top of bedrock. The age and character of the basal till are unknown. Small tongues of this advancing lobe extended into the cols at Stony Clove, Deep Notch and Grand Gorge, and ice scour and meltwater drainage through these cols served to excavate them to an elevation close to their present thresholds. Lingering masses of ice were present in the upper reaches of the tributaries south of each clove, for the outwash there is of ice-contact character. Recession of the ice block toward Hunter permitted the Red Kill to build a small delta with a summit elevation of 1890 feet. The present threshold elevation of Stony Clove is 2070 feet. The bottom of Stony Clove appears to have been elevated by landslide debris accumulation. Even if one allows for a 200 foot thickness of such a debris fill, it does not seem likely that a glacial lake at the elevation of the delta summit would have drained southward through Stony Clove. Ice-contact deposits lying between 1800 and 2000 feet occur one-half mile south of Devil's Tombstone (Stony Clove). Dissection of these ice-contact deposits is minor, easily accomplished by the steep tributary streams of upper Stony Clove Creek, several of which have built small, steep alluvial fans into the valley. There is little to suggest erosion here by lake spillway waters of any lengthy duration. At Haines Falls, the low point in the Kaaters Kill-Schoharie divide lies presently at 1940 feet with no indication there of an outlet channel draining eastward into the Kaaters Kill.

The Tannersville ice was apparently drained to the west through Grand Gorge (1620'+) at a time when ice still protruded through that col. This condition would allow deltas and kame terraces inferior to the Stony Clove (2070') and Deep Notch (1900')thresholds to form. Ice-contact deposits at Grand Gorge are missing, suggesting free drainage into the Delaware. This ice mass downwasted with abundant recessional deposits having decreasing elevations to the west. Left behind were detached ice blocks which served as local drainage controls for kame terraces and deltas lying between 1600 and 1400 feet in the valleys of the Schoharie and Batavia Kill. Minor readvances of the ice are indicated at Mosquito Point and Lexington where deformed sand and clay are overridden by till. How much of the Schoharie was deglaciated during this time is not presently known. Gravel underlying Lake Schoharie clay at Prattsville might suggest a free drainage northward perhaps with exit at the Franklinton col at an elevation of about 1100 feet.

Prattsville Margin

A second readvance of the Schoharie Lobe is proposed to have extended south to near Prattsville, with side lobes reaching to Grand Gorge and into the valleys of the Manor Kill and Platter Kill. Restoration of Lake Schoharie took place during this readvance where clays north of Prattsville with a summit of 1400 feet were deposited in lake water with at least this level. Build-up of the Hudson Lobe simutaneously sent tongues of active ice through the larger cols along the Northeast Escarpment. Two of these tongues at Broome Center and between High Knob and Steenberg Mt. (Stop 7) served to supply meltwater for extentive outwash and delta systems in the upper Platter Kill and Manor Kill valleys. A small lake formed between Manorkill and Conesville where a delta complex was built into each end. (Stops 6 and 8) Recession of this margin down the Platter Kill Valley produced a series of progressively lowered outwash and delta deposits between the Broome Center col and Gilboa (Stop 5). The upper Platter Kill and Manor Kill valleys record active ice thrusting over the divide from the east in the form of convex-westward moraine loops, mapped by Rich (1935). The contemporanity of these lobes with ice lying in the Gilboa-Conesville area is indicated by the abundance of coarse gravel and sand laid against ice (Stop 5) which fills the upper reaches of each valley. These deposits are too extensive to have formed by normal processes of precipitation and erosion, and must have been transported by actively moving meltwater. The Grand Gorge col served as an exit for this glacial drainage with deltas in the Manor Kill and Platter Kill between 1580' and 1620' suggesting the final stage in the downcutting of the Grand Gorge col. In the western Schoharie Basin presently drained by the Cobles Kill, the ice margin at this time rested against the extension of the Northeast Escarpment from Summit through the West Richmondville col and then into South Valley and Cherry Valley. Stages in the recession of the Schoharie Lobe are indicated by the ice-contact clay deposits lying in the valley bottom between Gilboa and Breakabeen (Stop 9). When the Hudson lobe had deteriorated so as to no longer contribute meltwater to the 1940-foot outlet channel west of Broome Center, a series of progressively lower kame deltas and eskers were formed in the Keyser Kill Valley. These range in summit elevation from slightly under 1900 feet down to 1400 feet. A sag south of the summit of Brown Mountain permitted meltwater to issue across the divide there and form a small delta at 1620 feet overlooking the Schoharie Valley at North Blenheim. Rich (1935) considered this delta to reflect the northernmost extension of his Lake Schoharie comparable in elevation to a 1620 foot temporary sill at the Grand Gorge col. The present writer prefers to substitute stagnating ice for much of the Lake Schoharie waters at this time, with the drainage from the Conesville-Manor Kill area, the Platter Kill and from the Keyser Kill extending over ice to exit for a time at Grand Gorge. Deterioration of the ice from this margin might expose again the col at Franklinton which could serve upon deglaciation of much of the upper Cats Kill to drain Lake Schoharie waters, but at a time while ice still lingered as detached blocks in the area between Breakabeen and Gilboa. Recession of this ice margin took place to at least the latitude of Middleburg in order to permit the accumulation of red clays in a lake between Middleburg and the Franklinton outlet. Deglaciation of the Cobles Kill Basin also permitted lacustrine conditions to develop in the basin now occupied by West Creek, a tributary to the Cobles Kill, between Seward and Hyndsville.

Middleburg Margin

A third readvance of Schoharie ice permitted the construction at the limit of the advance of the major portion of the Franklinton morainal plug (Stop 2) in addition to other end morainal features in the valleys of Little Schoharie Creek at Huntersland, the main Schoharie near Breakabeen, and the Cobles Kill near Richmondville. The lacustrines near Hyndsville were overridden with thin clay-rich till overlying clay with a terminus for the advance suggested by the valley-choking moraine west of Seward at the foot of the escarpment. In the easternmost Schoharie Basin at the head of Swiss Kill a kame moraine near Rensselaerville may correlate with this advance. Remnants of ice probably occupied parts of the upper Cats Kill establishing local lacustrine conditions adequate to maintain a kame delta at the Franklinton moraine with an elevation of 1280 feet (Stop 3). Recession of the ice from the moraine (Stop 1), and deterioration of the ice blocks in the Cats Kill permitted the final downcutting phase of the Franklinton col (Stop 4) with a threshold at present lying at 1180'. Ice override in the tributary valleys northeast of Peterburg Mountain is indicated by thin till overlying sand, clay and older till in well records. In the Cobles Kill Valley slight smoothing of kame terraces with occasional veneers of thin till are found southeast of the Cobles Kill between Mineral Springs and Richmondville. Inwash ice-contact deposits of gravel are found which correlate at near 1000 feet with recession of the ice from the Middleburg area and esker-like masses of clay south of Vroman's Nose suggest the lingering of ice in the valley bottom after some drainage of the lake waters from the 1180' outlet at Franklinton. If the clays south of Vroman's Nose with a summit elevation of 800 feet date from the time of outlet of Lake Schoharie at 1180' at Franklinton then a water depth of something under 400 feet would have been present at the time of their accumulation. This is a more desirable alternative to the 800-foot depth required by Rich's hypothesis. The compound character of the Franklinton morainal plug (gravel and clay overlying till) and the availability of a deglaciated Cats Kill Valley provides an alternative to the concept of large lake bodies which are required to overflow at high levels through the southern divide. It is clear that a single recessional deglaciation of an ice margin is inadequate to explain the complicated subsurface stratigraphy and the requirements of drainage of Lake Schoharie during the deglacial episode.

The recession of the Middleburg ice was accompanied by falling Lake Schoharie waters, but the point of exit from the Schoharie Basin for these waters is not clear. Upon exposure of the Helderberg limestone karst terrane leakage possibly took place through underground cavern systems. Another possibility might involve an exposure of an outlet near Delanson permitting overflow for a while at near 800 feet. Water well records for the Cobles Kill Valley in the vicinity of Howe Caverns suggest that the lower Cobles Kill was always occupied by either ice or lake water. Minor recessional ice-contact features are also found on the uplands east of Schoharie where several small esker segments trend down the hillslopes.

Yosts Margin

A final readvance of the Mohawk Lobe involved the lower Schoharie Basin, where a valley-choking moraine is now located just east of the junction of Fox and Schoharie Creeks. In the lower Cobles Kill Valley above Central Bridge a similar valley choking morainal system defended a small lake in which clays accumulated to 750 feet. Similar small, local lakes appear to have been imponded along the escarpment overlooking the lower Schoharie between Esperance and Glen. Yatsevitch (1969) has referred to a readvance through the Mohawk which terminated at the Noses on the Mohawk River. This readvance produced till which overlies stratified sand and gravel at Tribes Hill (Stop 15), and also produced an end moraine system running north and south along the Noses Escarpment. He calls this the Yosts Readvance. If earlier Lake Schoharie waters ever exited from the basin at Delanson, the record of these spillways was obliterated by this final readvance, unless the now-beheaded Bozen Kill at Duane records this earlier exit. A slight recession permitted the development of two outlet channels, which have been previously referred to as the Delanson outlet or the Delanson River (Fairchild, 1912). More complete stagnation of this final ice mass exposed the till and bedrock sill for the final controls of Lake Schoharie at Esperance (Stop 11). Recession through the lower Schoharie has left a somewhat fragmented record of kame terraces and gravel outwash systems between Burtonsville and Fort Hunter.

LAKE AMSTERDAM Final disappearance of ice from the Mohawk permitted the development of Lake Amsterdam with a water elevation of 420 to 450 feet. In its early stages Lake Amsterdam was blocked on the east by a narrow ice lobe as evidenced by concordant summits of deltas at Cranesville and Hoffmans. Final disappearance of the ice appears to have exposed a sill developed on till and limestone at Cranesville along the Cranesville fault. The duration of Lake Amsterdam with this spillway is not known but meltwaters from Port Huron ice in the western Mohawk or Lake Iroquois discharge may have finally destroyed the sill.

Stages in the downcutting of the Esperance sill for the final stage of Lake Schoharie are indicated by a series of terraces one-half mile north of Esperance. The upper, oldest sill level is indicated by a terrace at 680', which stands at the narrowest part of the present Schoharie gorge (Stop 11). Successively lower terraces are found at 640', 620', with modern terrace remnants at 580' and 560'. The northern limit of Lake Schoharie clays is found immediately south of the sill, with summit elevations near 650 feet. Clearly the sills at Esperance served to retain Lake Schoharie waters after the disappearance of ice from the lower Schoharie Valley, and it is probable that Lakes Schoharie and Amsterdam were in part contemporaneous.

OLDER GLACIATIONS

Several of the trip stops are at localities where till is found overlying stratified sediments. In these cases the occurrence can perhaps be best explained by minor readvances of Late Cary(?) ice which left a relatively thin record of its activity.

Several dozen water well logs obtained throughout the Schoharie Basin also show good evidence for multiple glaciation and in some logs the till portions each exceed 50-100 feet in thickness. Of particular interest are the well records from two thick-drift areas, one of which extends along Route 20 between Esperance and Sloansville, diagrammed in Figure 4. The second area extends from Central Bridge to Cobleskill and reflects a similar valley history. In many of the wells shown

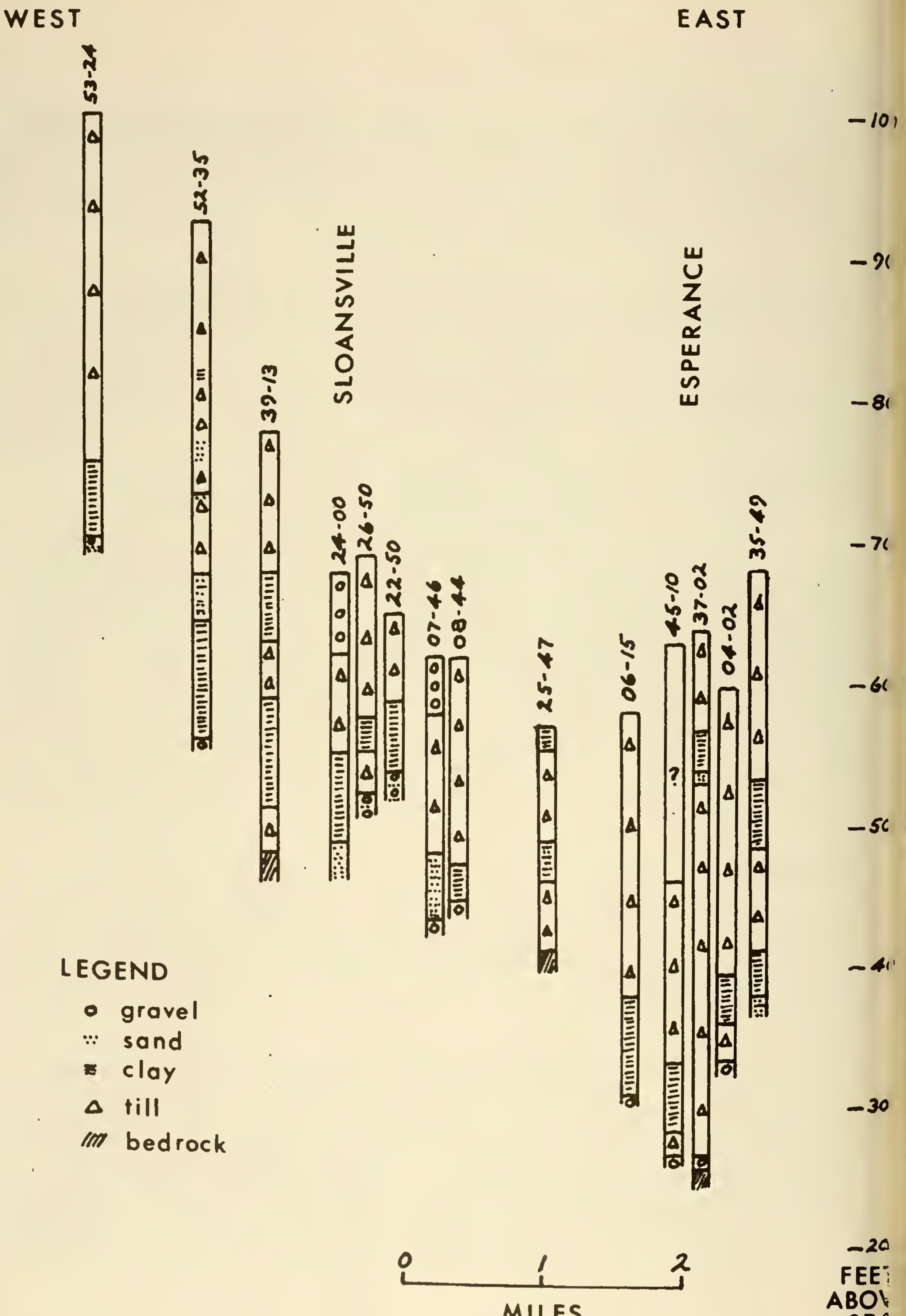

FIGURE 4. Water well logs indicating multiple glaciation in the Sloansville-Esperance area (Esperance quadrangle).

on Figure 4, the drift thickness exceeds 300 feet and in one, three glacial-deglacial cycles are represented. Well 39-13 shows three tills separated by two clay intervals. At least two cycles are indicated in several of the others.

The most important concern in correlating these records is in assigning correctly an age for the upper till. It is not clear at this writing whether the upper till represents all Late Wisconsin glaciation in this area deposited by ice which overtopped the Catskills, or whether it includes or is composed completely of till of some of the later readvances.

At Tahawas in the eastern Adirondacks, Muller (1965) and Craft (1969) have described a multiple till section which includes wood-bearing sediment with a radiocarbon date of > 40,000 years. If the upper till in these wells and at Tahawas represents the last major glaciation in southeastern New York, then a more ancient age for much of the drift in the Schoharie Basin is indicated.

GLACIAL HISTORY

Extinction of Lake Amsterdam. Mohawk delta at 340' into Lake Albany

Port Huron advance in western Mohawk Basin

Lake Amsterdam

Esperance sills

Delanson channels

Yosts Readvance

Franklinton channel

Middleburg Readvance

Grand Gorge outlet

Prattsville Readvance

Deglaciation of High Catskills: local alpine glaciers

Tannersville Readvance

Deglaciation, unknown extent

Main Cary (?) or older (?) advance, overtopping Catskills

Deglaciation with clay and sand deposition in lakes. (Pre-Cary (?) age). Free drainage of basin to Mohawk and Hudson?

Glaciation with deposition of lacustrine clay in lakes blocked by advancing ice. Override and till deposition. Extent and age of glaciation unknown.

Deposition of black sand (Braymanville aquifer) in Cobles Kill Valley

Preglacial northward drainage of Schoharie Valley, course between village of Schoharie and the Mohawk lying about a mile east of the present course.

SELECTED EXPOSURES
and
TRIP STOPS

Stop 1 Middleburg quad. 42 35 06 N 74 19 05 W 740' el. Small pit in esker segment behind gas station on Rt. 145 one mile southeast of Middleburg. Exposes 1-3' pebble gravel, overlying 15' sand with deformed red clay beds, overlying faulted and collapsed pebble sand. Formed upon withdrawal of ice margin from Franklinton morainal plug. Good view of plug to south and of ice-scoured Vroman's Nose and of alluvial fan of Little Schoharie Creek to west.

Stop 2 Middleburg quad. 42 34 29 N 74 18 36 W 970' el. Small exposure of till with subrounded boulders of local siltstone and a few clumps of red clay in matrix of red clay. May be till formed by Middleburg readvance over lacustrine sediments. Smooth slopes leading up to Franklinton plug, and ice-contact clays visible to south.

Stop 3 Middleburg quad. 42 33 27 N 74 18 55 W 1260' el. Lobate kame delta atop Franklinton moraine. Pit exposes boulder gravel with south dip. Abundance of subrounded local lithologies with few carbonates. May represent early exit of waters to Cats Kill while receding ice choked the valley, or may represent terminus of Middleburg readvance over dead ice and lacustrines in main Schoharie Valley to north. This deposit is graded to 1240' terrace remnants along Rt. 145 one mile north of Franklinton.

Stop 4 Middleburg quad. 42 32 50 N 74 18 08 W 1230' el. View of Franklinton outlet channel at 1180' and upper 1230' outwash terrace remnant at the present Schoharie-Cats Kill divide. For a mile to the south, small fans choke the channel and a swamp has formed.

Stop 5 Gilboa quad. 42 25 04 N 74 24 54 W 1620' el. Pit 2 miles NE of Gilboa exposes over 30' of well-rounded, SW-dipping lobate cobble gravel and sand laid against ice on downstream side. Supply of outwash was from ice lobe extending into head of Platter Kill Valley at Broome Ctr. 2 miles to the NE. Normal faults on SW end of pit. This locality is cited by Rich (1935) as evidence for upper 1620' phase of Lake Schoharie, with spillway at Grand Gorge.

Stop 6 Livingstonville quad. 42 22 51 N 74 21 41 W 1560' el. Conesville choker moraine. View to east toward Manor Kill and Steenburg Mt. col through which ice readvanced westward. Pit at 1560' exposes east-dipping pebble gravel and sand. Small lake basin lies to east.

Stop 7 Livingstonville quad. 42 23 43 N 74 18 55 W 1500' el. Morainal loop terrane NE of Manor Kill.

Stop 8 Livingstonville quad. 42 23 12 N 74 19 30 W 1500' el. Manor Kill delta. Pit exposes lobate pebble gravel and sand, dipping west; derived from ice projecting through Steenburg Mt. col.

Stop 9 Breakabeen quad. 42 30 23 N 74 25 23 W 750' el. Roadcut one mile south of Breakabeen on Rt. 30 in deformed interbedded red and gray clay. Southern end of a two-mile-long mass of kame-terrace clay.

Stop 10 Gallupville quad. 42 40 26 N 74 14 55 W 760' el. Pit along Rt. 43 one mile west of Gallupville exposes 10 feet of oxidized till overlying 20'+ of tightly cemented kame terrace cobble gravel.

Stop 11 Esperance quad. 42 46 28 N 74 15 25 W 680' el. Terrace remnant at 680' one mile north of Esperance represents highest Esperance sill for late Lake Schoharie. View to east of narrowest part of Schoharie gorge, cut in till. Between this point and Esperance, road drops over lower terraces lying at 670', 640', 620', and 580'.

Stop 12 Tribes Hill quad. 42 55 15 N 74 19 45 W 420' el. Landslide into Auriesville Creek one mile southwest of Auriesville exposes in the slide scar 2-3' silt and pebble gravel overlying 20' varved brown clay overlying 12' compact calcareous dark gray till. Base of till rests on a few inches of exposed yellow sand and pebble gravel.

Stop 13 Tribes Hill quad. 42 56 14 N 74 16 45 W 280' el. A 20-foot stream bank cut at Rt. 5S bridge over Schoharie Creek exposes (poorly at this writing) cobble gravel and sand overlying dark gray compact till, overlying interbedded till and mess-bedded sand and clay in 2-3" beds.

Stop 14 Tribes Hill quad 42 57 25 N 74 16 33 W 340' el. Roadcut along new Rt. 5 one-half mile east of Tribes Hill exposes 15-20' dark gray till overlying 15' rhythmic sand and dark gray clay over mess-bedded clay and clean sand. A few stones are found in the clay which at the base of the section rests on limestone with S75W striations.

Stop 15 Tribes Hill quad 42 57 50 N 74 15 12 W 500' el. Sand pit one mile northwest Ft. Johnson exposes 10-15' light brown till overlying 45'+ quartz sand with some indurated layers. One of the type exposures of till assigned to the Yosts readvance. Hill in which pit is dug resembles a drumlin nearly a mile long, oriented east-west.

Stop 16 Amsterdam quad. 42 53 45 N 74 09 59 W 620' el. Roadcut where Rt. 160 crosses Terwilleger Creek, exposes 10' brown till oxidized to base, and leached 3 ', overlying one foot of irregular 1-2" beds of gray till and sand, overlying one foot of sand and thin silt beds, overlying 40'+ gray unweathered till. Resembles the exposure at Stop 13.

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Trip 6

PALEOZOIC ROCKS IN WASHINGTON COUNTY, N.Y., WEST OF THE TACONIC KLIPPE

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INTRODUCTION

Trip 6 is a pendant to Trip 7, and the introduction to that trip gives the regional setting. The first stop of Trip 6 illustrates the geology at the west edge of the Taconic klippe; the second and fourth the stratigraphy of a western part of the carbonate bank (the eastern part is seen on Trip 7 in the Vermont marble valley); and the third some interesting structure in the carbonate rocks.

The detailed stratigraphic subdivision of the western carbonate shelf section is based mainly on relatively subtle differences in the carbonate rocks and their content of quartz sand and silt and on the rather sparse fossils (fossils are common only in the Middle Ordovician Glens Falls Limestone). It has been developed over a long period through the work of Brainerd and Seely (1890), Rodgers (1937), Wheeler (1942), Cady (1945), Rodgers (ms.), Welby (1961), Flower (1964)1/, Johnson and Welby (1966), and Fisher (ms.), not without considerable terminological confusion and controversy. A simplified section, applicable in this area, is given below (mainly from Johnson and Welby and from Fisher):

Middle Ordovician (Mohawkian)	Average thickness (feet)
Canajoharie Shale ("Snake Hill Shale")	600+
Black and gray shale, partly silty and sandy, partly calcareous, to east considerably deformed and cut by slaty cleavage	
Glens Falls Limestone	100?
Medium to dark gray (weathers dark gray) argillaceous to shaly thin-bedded fine-grained limestone, with beds of fossil debris	
Orwell Limestone	60
Light to medium gray (weathers light gray to nearly white) relatively pure rather massive very fine-grained limestone	
Disconformity; cuts out Chazyan strata and locally Providence Island Dolostone and upper Bascom Formation	

1/ The section given by Rodgers from Flower (Billings, Rodgers, and Thompson, 1952, p. 34-36, Table 2, Column 1) was based on an earlier version of Flower's article, which was supposed to be published ahead of the guidebook article, but Flower withdrew the manuscript before publication rather than permit the deletion of some doggerel verse (Flower, 1964, p. 161); thus the names as given by Rodgers are nude and <u>should never be cited as if properly published</u>.

Lower Ordovician (Canadian)	Average thickness (feet)
Providence Island (or Bridport) Dolostone (Upper Canadian) Light gray (weathers lighter gray or faintly yellow) mostly fine-grained well bedded to thick-bedded rather uniform dolostone; little quartz sand or silt	200
Bascom Formation, commonly divisible into Fort Cassin Formation (Upper Canadian) and Fort Ann Formation of Flower, 1964 (Middle Canadian) Highly varied alternation of medium to dark gray banded and laminated limestone, well bedded to massive dolostone, and dolomitic and calcareous sandstone; much sand and silt	200
Cutting Formation (Lower Canadian) (Great Meadows Formation of Flower, 1964, who considers that the rocks here do not correlate with the Cutting Formation of the Shoreham, Vermont, section) Light to dark gray (weathers medium to dark gray) commonly mottled fine- to medium-grained well bedded dolostone, with some quartz silt and considerable dark gray to black chert, especially in lower half; at top, lenses of very light-weathering pure light gray fossiliferous fine-grained limestone, upon which a karst surface is locally developed (as at Stop 2); at base, persistent member of very light gray slightly bluish dolomitic and calcareous fine-grained laminated sandstone or siltstone with conspicuous and characteristic cross-bedding, commonly containing fossil trails	200
Lower Ordovician and Upper Cambrian	
Whitehall Formation (of Rodgers, not Flower) Light to medium gray (generally weathers light gray) medium- to coarse-grained mostly poorly bedded relatively pure dolostone; some dark gray chert in lower part; lenses and persistent layers of pure light gray limestone (some weathers almost white). Persistent limestone unit near base (up to 75 feet thick at Stop 4) carries Trempeleau trilobites; limestone lenses near top carry Lower Canadian nautiloids	225
Upper Cambrian	
Ticonderoga Formation (see Rodgers *in* Welby, 1961, p. 232-234) Mainly dark gray fine- to medium-grained fairly well bedded dolostone; some layers contain much quartz sand and a few are dolomitic sandstone or quartzite; dark chert in upper part. Beds of *Cryptozoon*	300
Potsdam Sandstone Vitreous medium- to coarse-grained well bedded quartzitic sandstone or quartzite, commonly tan, cream, or pink, but locally almost black; rare beds of dolomitic sandstone or sandy dolostone. Carries Franconian and Dresbach trilobites. Rests unconformably on Precambrian gneiss	200

STOP 1, BALD MOUNTAIN

Schuylerville 15' quadrangle (Fig. 6-1 - area will straddle border of two eastern 7½' quadrangles when they are published); town of Greenwich, Washington Co., N.Y., 2 miles north of village of Middle Falls.

To reach stop, turn west off Route 40 at road junction 1 mile north of northeastern intersection of Routes 40 and 29 (northeast of Middle Falls) and drive northwest for 1.1 mile, passing first road intersection (at 0.7 mile), and park north of second intersection. Quarries are at southwest foot of Bald Mountain, which rises to northeast of parking spot, and other carbonate rocks are exposed in flat area between quarries and road, also on west side of road north of third intersection, and from first intersection south to and beyond Middle Falls.

Bald Mountain is a famous locality and has been subject to an extraordinary variety of interpretations over the years since Ebenezer Emmons first studied the locality and Asa Fitch found nearby the first "Primordial," i.e. Lower Cambrian, fossils known in North America. The mountain itself is formed of Cambrian slate, etc., of the Taconic sequence (Bomoseen and West Castleton Formations, according to Platt, ms., 1960), whereas the glacial lake plain west of the mountain is underlain by Middle Ordovician black shale ("Snake Hill Shale") and graywacke (Austin Glen Graywacke), presumed to belong at the top of the standard carbonate or shelf sequence of the region. In between, carbonate rocks belonging to the latter sequence crop out, not only in the two large quarries excavated at the west and southwest foot of the mountain but also in a belt extending from northwest of the quarries southward for nearly 5 miles past the village of Middle Falls. Black shale is also present within this belt of carbonate outcrop, and close to the carbonate rocks it contains fragments of them. The relations are most clearly seen in the two quarries (especially around the north end of each) and in their vicinity.

Everyone from Emmons' day has agreed that two quite different series of rocks are involved, what are called above the Taconic and carbonate sequences. Emmons (1844, p. 45; 1846, p. 89) considered that the carbonate rocks, which he agreed were part of the "New York System," i.e. the standard Paleozoic section of New York State west of the Hudson River, rest unconformably on the slate, which he assigned to his Taconic System; thus the locality provided one of the key arguments for his contention that the Taconic System was entirely older than the New York System (which began with the Potsdam Sandstone, now called Upper Cambrian). Barrande's identification of Fitch's fossils as "Primordial" seemed to confirm Emmons' age assignments. When Walcott studied the area, in the hope of resolving the bitter Taconic controversy that had ensued, he (1888, p. 317) recognized the contact as a thrust fault, bringing the older slate over the younger shale but not precluding contemporaneity of other parts of the two sequences; he concluded that the carbonate rocks at Bald Mountain form an ordinary thrust slice along the fault. Ruedemann

(Cushing and Ruedemann, 1914, p. 108-111, also accompanying plate) agreed, and he called the shale interleaved with and containing fragments of the carbonate rocks a mylonite developed along the fault. Rodgers (Billings, Rodgers, and Thompson, 1952, p. 49), impressed with the sedimentary appearance of the carbonate fragments in the shale, considered them exotic pebbles, and he further suggested that the two masses of limestone in the two quarries are simply mammoth boulders in the shale, but he accepted Walcott's view that the rest of the carbonate rock in the area is a thrust slice. Sanders, Platt, and Powers (1961, and personal communication) concluded that the pebbly slate is a basal deposit lying unconformably upon the quarry limestone and that all the carbonate rocks are autochthonous, brought up from below along a normal fault that happens to coincide areally with Walcott's thrust fault, whose existence they accepted. Recent field work by Bird and Rodgers and by Fisher and Davis (see Johnson and Welby, 1966, p. 62-64) indicates however that <u>all</u> the carbonate rocks are in the form of blocks in the black shale, that blocks of Taconic sequence rocks are mixed in, and that similar blocks, though generally less well exposed and less spectacular, can be found in a belt extending from west of Brandon, Vermont (where they include the Forbes Hill Conglomerate of Zen, 1967, p. 38, and other papers) to Newburgh, New York (Fisher, unpublished data), all along the western margin of the main Taconic slate mass or klippe, which now is itself interpreted as one or several immense blocks imbedded in the same Ordovician shale when it was still soft mud (at least the western Giddings Brook Slice is so interpreted; see Zen, 1967). One argument for this interpretation is that the carbonate rocks here, though unmetamorphosed, are stratigraphically more like those of the Vermont marble valley than of the western carbonate belt in the Champlain Valley and around the Adirondacks (e.g., Beldens instead of Providence Island or Bridport facies at the top of the Lower Ordovician).

We will visit the quarries, especially the north end of the southern quarry, after which visitors may wish to climb the mountain to the Taconic slate or visit the less well exposed blocks of carbonate (and other) rocks to the northwest and southwest.

Figure 6-1. Geologic map around Bald Mountain, Greenwich, N.Y.

Base from Schuylerville 15' quadrangle, enlarged to 1/24,000. Geology from reconnaissance notes by John M. Bird, John Rodgers, Donald W. Fisher, and James F. Davis.


Quarries
Middle Falls
Greenwich
(BM 360)
Cement
SEQUENCE
Ocl
40
29
350
375
335
336
500
B M
MAIN

STOP 2, SMITH BASIN SECTION

Hartford 7½' quadrangle (Fig. 6-2); towns of Hartford and Kingsbury, Washington Co., N.Y., 1 to 2 miles east-northeast of village of Smith Basin.

To reach stop, take Route 149 to the Bush farm (Bushlea) 1½ miles east of Smith Basin (3½ miles east of intersection with Route 4, or 4½ miles west of intersection with Route 40 in Hartford, N.Y.), park in farmyard and ask permission (he is used to geologists). Main section lies west of farm in hills north of road.

We are here at the south end of the continuous belt of Cambrian and Ordovician carbonate rock of the Champlain Valley - the northern Appalachian equivalent of the Valley and Ridge province (it is however much narrower and shorter, and no rocks younger than Middle Ordovician are present). The entire section is laid out in a monocline dipping gently eastward (5° to 20°), except that the Providence Island Dolostone is missing here beneath the Middle Ordovician. The hills immediately west of the Bush farm are in the lower part of the Bascom Formation, and the section continues down westward in the pastures to the top of the Whitehall Formation. The basal sandstone member of the Cutting is particularly well displayed just beyond the abandoned road near the town line, about a quarter of a mile north of Route 149. The lower part of the section can be seen in the woods beyond, down to the top of the Potsdam Sandstone near the Champlain Canal half a mile farther on. In the opposite direction, the upper part of the Bascom Formation (Fort Cassin Formation) is exposed east of the road on the Fish farm (the next farm northeast), and the Orwell and Glens Falls Limestones are exposed along Route 149 about 0.8 mile northeast of the Bush farm, where cleavage-bedding relations are especially well shown.

The section, and indeed the entire carbonate belt, is cut off on the west and southwest by one of the major Adirondack normal faults, downthrown on the west. Beyond it lies only Middle Ordovician shale, except for the Middle Ordovician limestones exposed in quarries to the southwest across the Big Creek valley, either as separate thrust slices or as large exotic blocks in the shale; the base of the western slice or block is exposed in narrow cuts from the western quarry west to the road southeast from Smith Basin. Southeast of the Bush Farm, the Middle Ordovician shale extends to the western margin of the Taconic klippe, in the hills southeast of Hartford, 3 miles from the farm. The normal fault mentioned brings up the Precambrian on its east side southeast of Fort Ann, about 2 miles northwest of this stop; the Precambrian is well exposed at Battle Hill along Route 4 northeast of Fort Ann. The unconformable and slightly conglomeratic base of the Potsdam Sandstone is exposed in the roadcut next west of the railroad bridge on Route 4, but it is not easy to stop cars near it; splendid exposures of the unconformity can now be seen in fine new road cuts on Route 22, northwest of the village of Putnam Center, about 19 miles north of Whitehall or 30 miles north of Fort Ann.

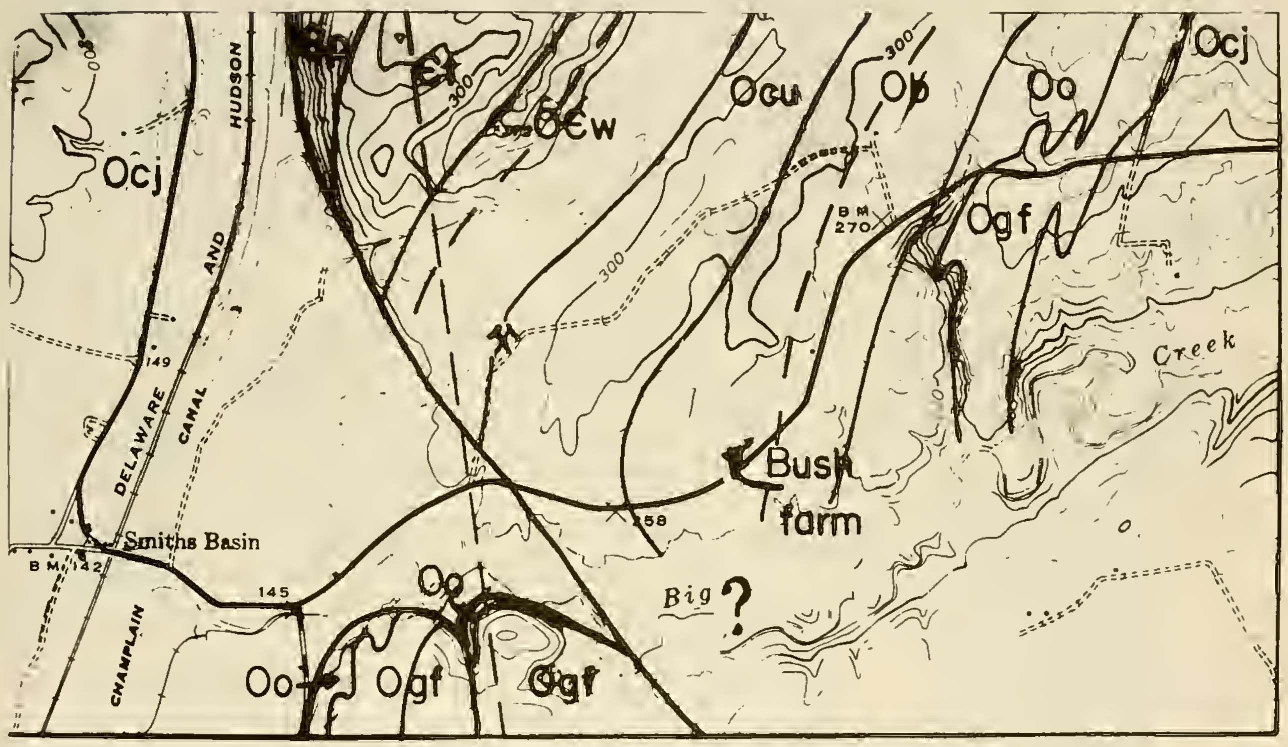

Figure 6-2. Geologic map around Smith Basin, Kingsbury and Hartford, N.Y. Basin from Hartford 7½' quadrangle, 1/24,000. Geology by Donald W. Fisher.

STOP 3, CHEVRON FOLDS ON TYLER FARM

Fort Ann 7½' quadrangle (Fig. 6-3); town of Fort Ann, Washington Co., N.Y., 2 miles southeast of village of Comstock.

To reach stop, turn east off Route 4 at Dewey Bridge 2½ miles northeast of Fort Ann (0.6 mile northeast of railroad bridge mentioned at end of description of Stop 2), drive east for 2.2 miles on Dewey Bridge Road to Tyler farm; ask permission (he also is used to geologists) and park either in farmyard or about two-tenths of a mile farther along the road, cross fence on north side of road (be careful with fence) and climb to top of first rise in pasture, on north side of which are chevron folds in Orwell and Glens Falls Limestones.

The belt of rocks exposed in the Smith Basin section (Stop 2), or at least its lower part, can be followed northward as far as Whitehall (Stop 4) and beyond, always dipping gently eastward though interrupted and slightly offset by a few east-west cross-faults. Beginning about 4 miles north of Route 149, however, it is bounded on the east by a thrust sheet in which the carbonate sequence is repeated, also dipping gently eastward but cut off in that direction by a large normal fault, downthrown on the east. The south end of the thrust sheet lies in the hills north and northeast of the Tyler farm, but it is abrupt and complex; instead of a simple rise of the thrust surface above the ground level or a straight-forward tear fault, one finds these very tight folds in the highest beds of the carbonate section, the folds plunging south <u>away</u> from the main body of the thrust sheet, as if it were disappearing underground. A small outlier, perhaps an isolated klippe of the same thrust sheet, overlaps the western section in the hills about 2 miles southwest of the Tyler farm, adding to the peculiarity of the structure in this area.

The west margin of the Taconic klippe is in the hills about 3 miles east-southeast of the Tyler farm, and carbonate blocks like those at Bald Mountain crop out in the shale just west of the klippe (in a good light, one or more can be seen from here).

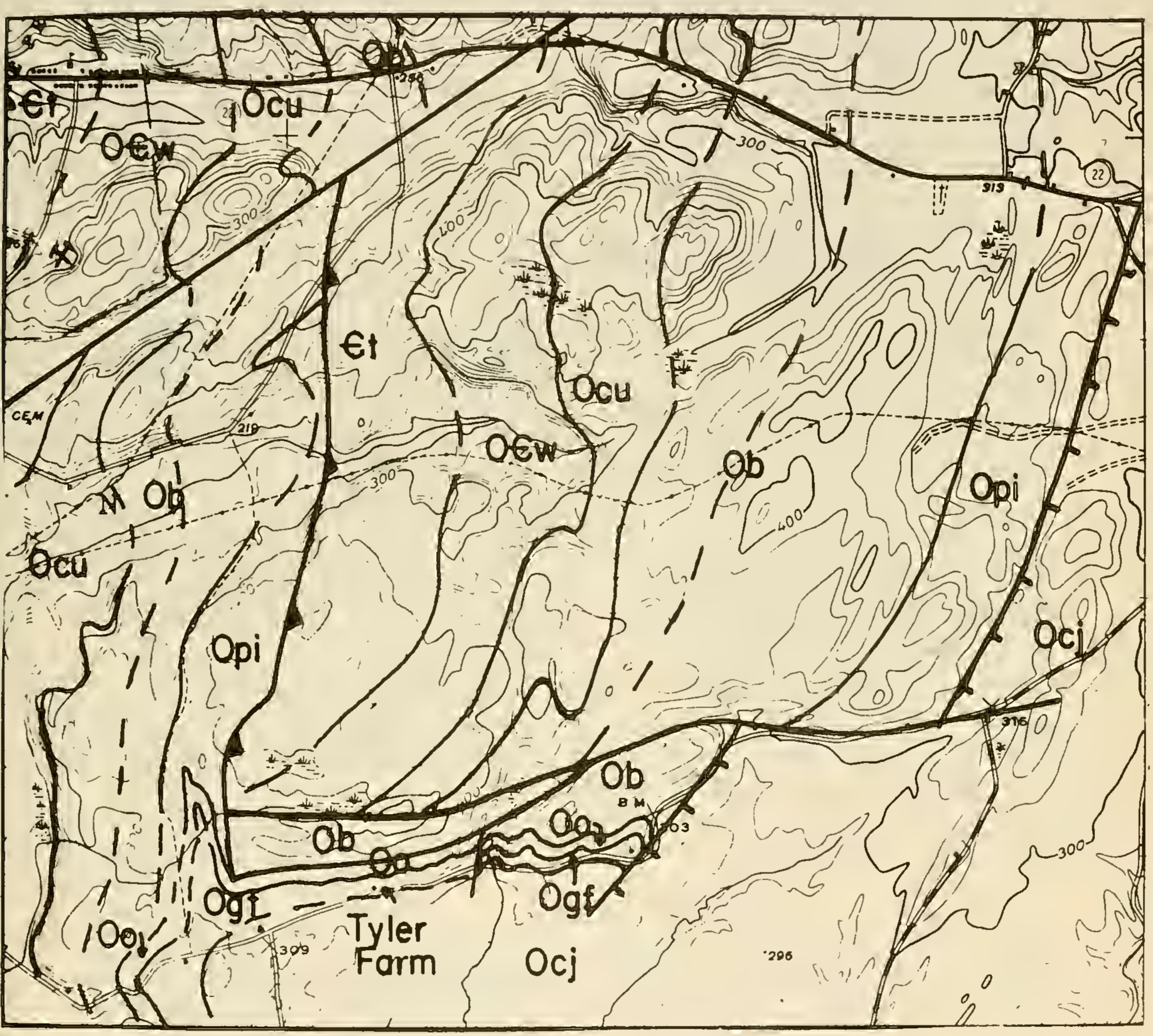

Figure 6-3. Geologic map around Tyler farm, Fort Ann, N. Y.

Base from Fort Ann 7½' quadrangle, 1/24,000. Geology by Donald W. Fisher.

STOP 4, UPPER CAMBRIAN LIMESTONE IN WHITEHALL FORMATION

Whitehall 7½' quadrangle (Fig. 6-4); town of Whitehall, Washington co., N.Y., 1 mile northeast of village of Whitehall.

To reach stop, turn east off Route 22 in north part of village of Whitehall (0.6 mile north of intersection with Route 4), cross canal and turn north (left), proceed 0.6 mile and turn north (left) again, proceed 0.6 mile and turn east (right), proceed 0.4 mile (on Town Route 10) to quarry on north (left) side of road.

In this area the Whitehall Formation seems to have a maximum amount of limestone, largely concentrated in a single unit, up to 75 feet thick, in the lower (Upper Cambrian) part of the formation (so-called "Hoyt member"). There are two kinds of limestone here. One is light gray lime-mudstone or calcilutite, weathering even lighter (ashy) gray and containing heads and biostromes of <u>Cryptozoon</u> (also some black and greenish chert). The other is darker gray darker weathering coarser limestone (lime-sandstone or calcarenite), with quartz sand in places, filling in around the heads and containing occasional Upper Cambrian (Trempeleau) trilobites.

The rocks here are in the same belt as Stop 2 and show the same gentle eastward dip (here about 10^{o}); the carbonate rocks are exposed in east-west blocks separated by relatively wide east-west valleys, apparently localized along belts of closely spaced joints or transverse faults with little displacement. Darker quartzose dolostone at the top of the Ticonderoga Formation is exposed below to the west, and light coarse-grained ("sugary") dolostone of the upper part of the Whitehall Formation ("Skene member") above to the east. Lower Canadian (lowest Ordovician) fossils are known in limestone layers in that part of the formation, but not near here. The type section of the Whitehall Formation is on the top and east slope of Skene Mountain, which rises directly out of the village of Whitehall about a mile southwest of Stop 4.

Those going north into Vermont may wish to make an additional stop at the new cuts on Vermont Route 22-A about 3 miles northwest of Fair Haven, where the shale next west of the Taconic klippe contains blocks of carbonate rocks like those near Bald Mountain (Stop 1).

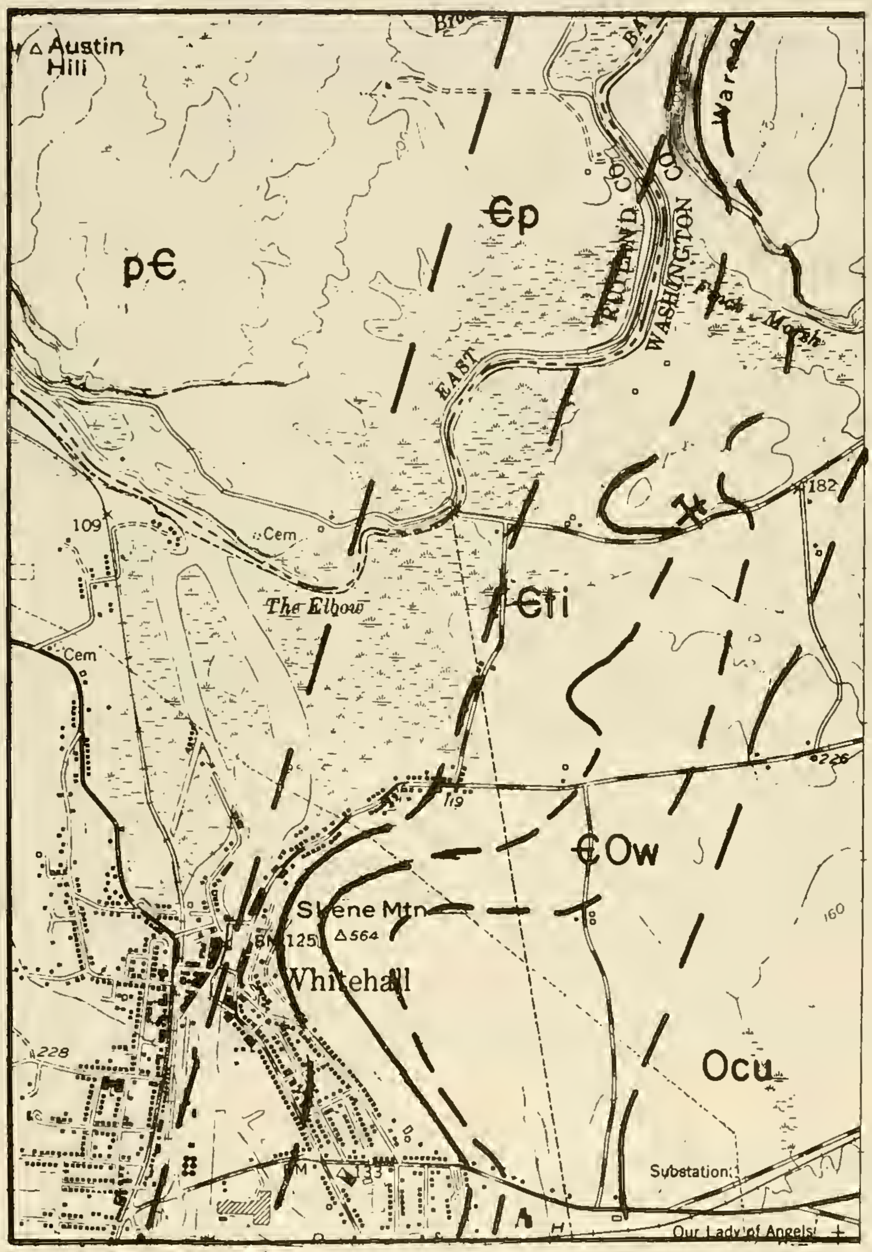

Figure 6-4. Geologic map around Whitehall, N. Y.

Base from Whitehall 7½' quadrangle, 1/24,000. Geology by Donald W. Fisher and John Rodgers.

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Trip 7

STRATIGRAPHY OF THE CHAMPLAIN VALLEY SEQUENCE IN RUTLAND COUNTY, VERMONT, AND THE TACONIC SEQUENCE IN NORTHERN WASHINGTON COUNTY, NEW YORK

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INTRODUCTION

The Taconic sequence, dominantly of Cambro-Ordovician shales with subordinate sandstones and limestones, extends some 150 miles from west-central Vermont southward along the Hudson Valley into Dutchess County, New York. It is juxtaposed with a contemporaneous sequence consisting essentially of sandstones and carbonates.

The Taconic area has been the subject of major controversies since 1842, when Emmons applied the name Taconic to certain rocks near Williamstown in western Massachusetts; he considered the Taconic to be older than what we now call the Ordovician. (See also discussion in Rodgers and Fisher, this volume, trip 6.) Emmons was later shown to have been partly correct when parts of his Taconic System yielded Cambrian fossils; other parts proved to be Ordovician.

Ulrich and Schuchert (1902) noted that the Early Paleozoic strata of eastern North America showed lithologically and paleontologically distinct sequences that have mutually exclusive geographic distributions. They proposed, following an idea earlier suggested by Logan (1866), that these sequences were deposited in distinct basins separated by a land barrier. Thus, in their view, the Taconic sequence was deposited in their proposed Levis channel and the contemporaneous and now juxtaposed sandstones and carbonates in their proposed Chazy basin, the two basins being separated by their proposed Quebec barrier. This idea carried the implication that the Taconic sequence had been thrust over the sandstones and carbonates.[1]

[1] Lochman (1956) reinterpreted the paleobiogeography of the Cambrian shelly faunas, pointing out that the occurrence of both Pacific and Atlantic province faunal elements in the *Elliptocephala asaphoides* fauna was inconsistent with the concept of a continuous land barrier (New Brunswick geanticline) acting as the biogeographic barrier between the two faunal provinces. She considered that mixing of faunal elements in the *E. asaphoides* fauna required deposition of the Taconic sequence in approximately its present geographic relationship to the sand-carbonate sequence. Furthermore, she developed the idea that the *E. asaphoides* fauna was indigenous to a deeper water site of mud deposition lying within an embayment in the sand-carbonate shelf, the Pacific province faunas being indigenous to this shelf.

In 1909, Ruedemann explicitly suggested that the Taconic rocks occupy a klippe; his suggestion was based on the contrast in lithofacies supported by the presence of certain thrust faults and tight folds. This started a second major controversy concerning the gross structural relationship of the Taconic sequence to the juxtaposed sandstone and carbonate sequence. A klippe hypothesis has, in general, been favored by geologists who have worked at the north end of the Taconics and opposed by those worked at the south end. Recent work by Zen (1967) has shown that the evidence is strongly in favor of a klippe hypothesis.

The contrast in lithofacies shown by these juxtaposed Cambro-Ordovician sequences is thus relevant to the historical development of the idea of the Taconic klippe. Furthermore, the stratigraphy has its own intrinsic interest, especially as the sections are among the best dated paleontologically in the northern Appalachians.

LITHOFACIES, TECTONIC ELEMENTS, AND SEDIMENTARY HISTORY

The western New England-eastern New York area included a number of tectonic elements during the Cambrian and Ordovician: (1) the craton, constituting the most westerly element and now represented in outcrop by the Grenville rocks of the Adirondacks, the Green Mountains, and smaller massifs; (2) a sand-carbonate shelf; and (3) a basin or eugeosyncline situated east and southeast of the shelf.

The rocks of the shelf are dominantly clean sandstones and carbonates forming interdigitating wedges; they are included in the Champlain Valley sequence. Generally, the carbonates thicken eastward and southeastward at the expense of the sandstones; the sandstones probably coalesce at depth westward and overlap progressively onto the craton. The Potsdam Sandstone, which can be seen in contact with the Precambrian of the Adirondacks, yields Dresbachian, Franconian, and Trempealeauian fossils (Fisher, 1956, p. 325, 339) and probably represents the present western edge of these coalesced sandstone wedges. This shelf extends for hundreds of miles along the Appalachians. Its width, estimated by Palmer (in press a) as originally of the order of 150 kms, has been reduced by later thrusting.

In west-central Vermont, the base of the sand-carbonate shelf sequence rests on the Dalton Formation consisting essentially of schistose graywacke and conglomerate with subordinate phyllites and impure dolomites. These may be regarded as basin deposits and they rest unconformably on the Precambrian of the Green Mountains and smaller massifs. The top of the sand-carbonate shelf sequence is marked by the contact with the Hortonville Formation consisting essentially of gray to black phyllite. This change, from carbonate to black shale sedimentation, is regional and almost synchronous from Quebec to Pennsylvania; in the vicinity of the Taconics, it represents a permanent change in sedimentation patterns (Zen, 1967, p. 10). Ordovician carbonate units, higher than those demonstrated in trip 7, are seen in the Whitehall-Fort Ann area (trip 6).

The basin[2] or eugeosynclinal sequence is now represented by a thick succession of schists and gneisses, with important graywackes and volcanics near the bottom and top, situated east of the Green Mountains. The dating of these strata is discussed by Cady (1960, p. 553-554), and Cady, Albee, and Chidester (1963, p. 7).

A relatively unmetamorphosed, westward-thrust, erosional remnant of this basin sequence is now preserved in the Taconic klippe. Emplacement of this klippe during the Taconic orogeny has been discussed by Zen (1967; 1968) and Cady (1968 b). The Taconic sequence is thinner than the thickest part of the eugeosynclinal sequence (Cady, 1967, p. 59, 63; 1968a, p. 155; 1968b, p. 570; Zen, 1967, p. 46).

Zen (1961; 1964), Theokritoff (1964) and earlier workers accepted an unconformity within the Taconic sequence between the West Castleton Formation, bearing Lower Cambrian fossils, and the Hatch Hill Formation, bearing latest Cambrian graptolites. This unconformity was postulated largely to account for the absence of demonstrated Middle Cambrian strata; it was supported by Dale's (1899, p. 291) report of local angular unconformities between these two formations. The subsequent discovery of Middle Cambrian fossils in Columbia County, New York (Rasetti and Bird, in Fisher, 1964; Rasetti, 1967; Bird and Rasetti, 1968) seriously weakened the earlier argument based on the supposed absence of Middle Cambrian fossils. Although Middle Cambrian fossils older than the Bathyuriscus-Elrathina zone are so far unknown from the Taconic sequence, the view that an attenuated section, rather than an unconformity, better explains the field data, now appears to be established in one form or another (Zen, 1967, p. 16-17, 47; Bird and Rasetti, 1968; Rasetti and Theokritoff, 1967, p. 192; Theokritoff, 1968, p. 19). Zen (1967, p. 47) and Rodgers (1968, p. 143) considered that the Taconic sequence was deposited in that part of the basin, at the foot of the shelf, that was starved of sediment. On the other hand, Cady (1967, p. 62-64; 1968a, p. 157; 1968b, p. 569-570; in Zen, 1967, p. 47) considered that the relative thinness of the Taconic sequence may be due to unconformable overlap and stratigraphic convergence related to the Vermont-Quebec geanticline, one of three "...relatively stable tracts of slowed deposition, nondeposition, and local erosion during episodic general uplifts of the northwestern half of the orthogeosynclinal belt, and of relatively mild deformation during subsequent orogenic folding" (Cady, 1968a, p. 152).

PALEOECOLOGY AND PALEOBIOGEOGRAPHY

The oldest known faunas in western New England and adjacent New York are the *Elliptocephala asaphoides* fauna of the Taconic sequence and the *Olenellus*-bearing faunas of the shelf sequence; older, pre-*Olenellus* faunas, known from California and Morocco, are absent. Hence, Theokritoff (1968, p. 9) suggested that the earliest Early Cambrian strata are represented in the area by unknown thicknesses of unfossiliferous strata conformably beneath the fossiliferous Cambrian. The base of the

[2] The term basin is used here to include sequences, such as the Taconic, that lack the volcanics characteristic of eugeosynclines but that are otherwise lithically and sequentially similar to the sequences with volcanics.

Cambrian cannot, therefore, be precisely fixed stratigraphically, but the presence of graywacke and volcanics (Pinnacle Formation with its Tibbit Hill volcanic member) in the lower part of the section suggests that a eugeosyncline was probably already established in the latest Precambrian.

Because Olenellus-bearing faunas are known from the Cheshire Quartzite, the lowest unit in the sand-carbonate shelf, it appears probable that the shelf was established somewhat later, during the Early Cambrian. The shelf persisted into the Middle Ordovician. Its sands were derived from the craton and probably represent marine reworking of the material of a huge sheet of arkosic sediment that must have been spread over the craton consequent to the erosion of the Grenville mountains; its carbonates probably originated in the shallow water over the shelf.

The occurrences of the Elliptocephala asaphoides fauna and the several other Taconic shelly faunas (Bird and Rasetti, 1968; Rasetti, 1966; 1967; Rasetti and Theokritoff, 1967; Theokritoff, 1959; 1964) are most numerous in limestone-conglomerates and bedded limestones enclosed in the shales of the Taconic sequence. It is hence inferred that the organisms were indigenous to sites of lime deposition on the shelf margin or slope and that they had been carried into the sites of dominantly mud deposition by means of slides and turbidity flows (Bird and Theokritoff, 1967; Theokritoff, 1964, p. 183-184; 1968, p. 18-19; see also Rodgers, 1968, p. 143). However, members of the E. asaphoides fauna are also known from shales of the West Castleton Formation at Judson Point, Columbia County, New York (Bird and Rasetti, 1968, p. 31) and at the type locality of E. asaphoides and Atops trilineatus (Emmons, 1844; 1846) three miles north of Bald Mountain in Washington County, New York. Bird and Rasetti (1968, p. 56) considered that the complete specimens of Serrodiscus speciosus found in the shales at Judson Point show no evidence of having been transported; it is possible that S. speciosus as well as other members of the E. asaphoides fauna may have had a limited adaptability to the basin environment. Rasetti (1967, p. 97) has pointed out that Atops trilineatus is known almost exclusively from the shales and extremely rarely from the limestone. Thus, A. trilineatus may be a species adapted to the basin rather than the shelf environment.

The age span of the Elliptocephala asaphoides fauna probably overlapped to a greater or lesser extent that of the Olenellus-bearing Pacific province faunas; hence the distinctions between the faunas represent biogeographic differences rather than chronological sequence, the Pacific province faunas occupying the inner margin of the shelf and the Taconic faunas, such as the E. asaphoides fauna, the outer shelf margin or slope (Theokritoff, 1968, p. 18-19).

Theokritoff (1968, p. 17) discussed the biogeographic affinities of the Elliptocephala asaphoides fauna and showed that although this fauna exhibits affinity with both Atlantic and Pacific provinces at the generic and especially the familial levels, it does not consist of a mixture of Atlantic and Pacific species. Hence it did not occupy a transitional environment between the two provinces lying within the range of adaptability of species otherwise characteristic of one or the other. He proposed to characterize such faunas as intermediate zone faunas, using the term zone in an ecological sense. The somewhat greater affinity expressed at the specific level with the Atlantic province was taken to suggest that there was a greater environmental affinity with that province than with the Pacific.

The overlying Paedeumias-Bonnia, Pagetia connexa, and Pagetides elegans faunules show somewhat stronger Pacific affinities (Rasetti and Theokritoff, 1967, p. 192-193).

Because the Olenellus-bearing Pacific province faunas occur in the sand-carbonate strata of the shelf and in shales that were deposited over the shelf, and furthermore, because the intermediate zone faunas also occur within the carbonates, the biogeographic barrier that separated them could not have been directly related to gross lithofacies. However, the geographic separation of the faunas, the Olenellus-bearing indigenous to the inner portion of the shelf and the intermediate zone to the outer margin or slope, suggests that the barrier may be related to the topography of the shelf and that it was the gross differences between the environments that constituted the barrier. A related factor may lie in the distribution of dolomite in the sand-carbonate shelf. Rodgers (1968, p. 142) stated that generally dolomites predominate to the west and northwest and limestone to the east and southeast, although their mutual facies boundary migrated in the course of time. Palmer (in press b) has presented evidence for an alternative interpretation with dolomites predominating in the central belt of the shelf. If so, and if the presence of dolomites is related to hypersaline conditions, it is possible that very shallow water conditions on top of the shelf may have been conducive to the development of a central hypersaline belt flanked by areas of normal salinity and limestone deposition to the northwest and southeast.

TACONIC SEQUENCE: FORMATIONS IN NORTHERN WASHINGTON COUNTY, NEW YORK

	Approximate Thickness In Feet
ORDOVICIAN	
Pawlet Formation: Graywackes and interbedded black shales. The graywacke is coarse, buff- to brown-weathering, very dark gray in fresh surfaces, and contains conspicuous quartz grains and locally fragments of greenish-gray slate.	100[3]
Berry (1962, p. 712-713) treated the Pawlet Formation as the Austin Glen Member of the Normanskill Formation; he (1962, p. 714) referred the graptolite faunas of the Pawlet Formation to the zone of *Climacograptus bicornis*.	
Indian River Slate: Red and blue-green slate, glazed in appearance, showing very little banding and with some cherts locally.	200
Berry (1962, p. 711) treated the Indian River Slate as a member of the Normanskill Formation; he (1961, p. 226; 1962, p. 711) correlated the Indian River Slate with the lower part of the Normanskill.	
Poultney Slate:	
Member C: Siliceous, dark-gray to black, banded argillite with local thin quartzites.	0-300
Member B: Light-buff- or white-weathering, bluish-gray or locally dark purplish-red, waxy looking, or cherty, layered argillites; also local developments of alternating finger-thick green and greenish-weathering black argillites and brown sandy beds; local thin partings of black shale. Toward the top, the argillite becomes more bluish and more waxy or cherty in appearance; there are small indigo-colored patches in the rock.	0-400
Member A: Dark-gray or black slates with local beds of white-weathering, dark-gray, sublithographic limestone and local lenses of limestone conglomerate containing pebbles of black chert.	0-100

[3] Thickness from Berry (1962, p. 713); all others from Theokritoff (1964).

Approximate Thickness
In Feet

The Poultney Slate ranges in age from Early Tremadoc to Middle Ordovician (Berry, 1961, p. 226)

CAMBRIAN

Hatch Hill Formation: Sooty, black, pyritic, rusty-weathering shales interbedded with rotten-weathering, bluish, dolomitic sandstones, locally cross-bedded and characteristically traversed by numerous quartz veins. — 0-350

Dendroid graptolites from the Hatch Hill Formation were considered by Berry (1959, p. 61; 1961, p. 224; in Bird *editor*, 1963, p. 25) to be Late Cambrian. As these graptolites were collected from near the top of the formation (Theokritoff, 1964, loc. 55-9), the base of the formation may be Franconian or even Dresbachian (Theokritoff, 1964, p. 179); Zen (1967, p. 54) considered the basal part of the Hatch Hill Formation to be Middle Cambrian.

West Castleton Formation: Dark-gray or black shales locally with several beds of fine-grained, dark-bluish-gray limestone from 6 inches to 1 foot thick or lenses of limestone conglomerate. — 0-150

The limestones are fossiliferous, especially in the western part of the area. They yield the late Early Cambrian *Elliptocephala asaphoides* fauna (for check lists, see Rasetti, 1967, p. 16 for trilobites; Walcott, 1912, for brachiopods; and Lochman, 1956, p. 1355-1356, for miscellaneous fossils), and in northern Washington County, three slightly younger but mutually approximately synchronous faunules: the *Paedeumias-Bonnia* faunule at one locality (Theokritoff, 1964, 58-11; see discussion in Theokritoff, 1964, p. 186; Rasetti and Theokritoff, 1967, p. 191); the *Pagetia connexa* faunule at three localities (USNM loc. 34; USNM loc. 38a; USNM loc. 43a; see discussion in Rasetti and Theokritoff, 1967, p. 191-192); and a faunule with *Pagetides elegans* and olenellid fragments at one locality (Rasetti, loc. cs-60; see discussion in Rasetti and Theokritoff, 1967, p. 192).

Mettawee Slate: Purple and green slates with local thin beds of limestone and limestone conglomerate near the top of the formation; lenses of quartzite occur locally. — 0-400

The boulders in the limestone conglomerates yield fossils of the late Early Cambrian *Elliptocephala asaphoides* fauna.

Approximate Thickness In Feet

<u>Bomoseen Graywacke</u>: Graywackes, typically dark olive-green and laminated, with subordinate green slates and quartzites. — 600+

Although the Bomoseen Graywacke has yielded no authenticated fossils, it is probably Early Cambrian because it lies conformably below strata containing late Early Cambrian fossils.

CHAMPLAIN VALLEY SEQUENCE: FORMATIONS IN THE RUTLAND AREA, RUTLAND COUNTY, VERMONT

MIDDLE ORDOVICIAN

Approximate Thickness In Feet

Ira (Hortonville) Formation: Gray to black phyllite; beds of blue or blue-gray limestone (Whipple Marble) in lower part contain Middle Ordovician fossils (Foerste, 1893; Thompson, 1967, p. 82-83; Zen, 1964, p. 39-42). Base of formation contains a variety of rock types, not all present in any one section. These include: blue limestone; dark gray dolomite or dolomite breccia; and a quartzite, locally with pebbles of blue quartz. — 500+

The Ira Formation cuts unconformably across all the older units of the Champlain Valley sequence (Thompson, 1967, p. 81)

Baker Brook Volcanics: Quartz-muscovite-biotite schist with feldspar augen, a greenstone schist containing a green actinolitic amphibolite; probably metamorphic derivatives of intermediate to mafic pyroclastics. — 200

The Baker Brook Volcanics rest unconformably on carbonates assigned to the Bascom Formation, to the Shelburne Marble, or to the Clarendon Springs Formation.

LOWER ORDOVICIAN

Bascom Formation: Thompson (1967, p. 79) mapped as Bascom Formation not only the Bascom Formation of Cady (1945) but also rocks like the Beldens Member of the Chipman Formation. The Cutting Dolomite of Cady (1945) passes laterally south of Brandon, Vermont, into rocks indistinguishable from the lower Bascom (for references, see Thompson, 1967, p. 79). In the Rutland area, rocks of 'Beldens' type overlie rocks of 'Bascom' type in areas of good outcrop and simple structure; elsewhere, the distinction could not be maintained and was therefore abandoned. — 350-450

Blue-gray and white calcite marble, locally with dolomite mottling and containing interbeds of gray, yellow, or orange-weathering dolomite.

Rocks on the west limb of the Middlebury synclinorium that are probably equivalent to the lower Bascom contain fossils indicating correlation with the upper Gasconade or lower Roubidoux of the Mississippi Valley (Cady, 1945; Welby, 1961; Flower, 1964).

LOWER ORDOVICIAN OR UPPER CAMBRIAN

	Approximate Thickness In Feet
Shelburne Marble: White calcitic marble with minor beds of light gray marble; lithologically similar to marble of Sutherland Falls Member of Clarendon Springs Formation but shows less dolomite 'curdling.' The Shelburne Marble contains fossils near Shoreham (Cady, 1945; Flower, 1964) correlative of the Gasconade of the Mississippi Valley, hence reference of the Shelburne Marble to a System, whether Cambrian or Ordovician, depends on the definition adopted for this Systemic boundary (see Fisher, 1962; Whittington and Williams, 1964). Welby (1961) correlated the Shelburne with the Whitehall of Rodgers at Whitehall (see Rodgers and Fisher, this volume, trip 6), and the Baldwin Corners of Flower (1964) of the Fort Ann area, New York.	250
Clarendon Springs Dolomite: Thompson (1967, p. 77) redefined the Clarendon Springs Dolomite of Keith, as mapped in the marble belt, to include two higher units previously mapped as the Sutherland Falls Marble and the Intermediate Dolomite, and both previously treated as members of the Shelburne Marble. This extended unit was named the Clarendon Springs Formation and includes the Clarendon Springs Dolomite, as formerly defined, as its lower member; the Sutherland Falls Member; and the former Intermediate Dolomite of the Shelburne as its upper member.	
Upper member: Light gray calcitic dolomite with abundant irregular masses of recrystallized chert.	150-200
Sutherland Falls Member: White calcite marble with minor beds of light gray marble; dolomite 'curdling' not uncommon.	50-100
Lower member: Light gray calcitic dolomite; lower part has beds, several inches or a foot thick, of cross-bedded sandy dolomite.	200-250

The lower member (formerly the Clarendon Springs Dolomite) yields faunas correlated with the *Hungaia magnifica* fauna which Whittington (1966, p. 701) considered to be Trempealeauian and also correlative of the early Tremadoc (for references, see Theokritoff, 1968, p. 15).

Stone and Dennis (1964, p. 53-54) reported an occurrence of *Ellesmeroceras* sp. considered by Flower to be lowest Ordovician; the stratigraphic position of the fossil locality is ambiguous and further work is needed to clarify it.

Approximate Thickness In Feet

Cady (1945) and Welby (1961) correlated the Clarendon Springs with the Ticonderoga Dolomite of Rodgers at Whitehall, New York (see Rodgers and Fisher, this volume, trip 6) and the Dewey Bridge and 'Whitehall' of Flower (1964) near Fort Ann, New York (see discussion in Thompson, 1967, p. 78).

UPPER CAMBRIAN

Danby Formation: Light gray calcitic dolomite interbedded with cross-bedded sandy dolomite or dolomitic quartzite and, in the lower part, beds of pure vitreous quartzite that may be five or six feet thick. — 50-150

Cady (1945, p. 530) has traced the Danby around the north end of the Middlebury synclinorium to outcrops at Shoreham, Vermont, virtually lithologically identical to the Potsdam Sandstone at Fort Ann and Whitehall, New York. Here, the Potsdam Sandstone contains Dresbachian and Franconian fossils (Flower, 1964). The Danby Formation is hence assigned to the Upper Cambrian.

MIDDLE CAMBRIAN

Winooski Dolomite: Light gray dolomite, typically weathering yellow or orange-red. Beds are commonly several inches thick separated by thin siliceous partings. Minor rock-types include dark blue-gray dolomite, rusty weathering dolomitic quartzite, and greenish schistose quartzite. — 300-400

No fossils have yet been found in the Winooski Dolomite or in the Rugg Brook Dolomite, its equivalent in northern Vermont. However, the Winooski Dolomite is assigned to the Middle Cambrian because the Rugg Brook overlies the Parker Slate which contains Middle Cambrian fossils in its upper portion, and also underlies the St. Albans Slate which contains Middle Cambrian fossils. (Shaw, 1966)

LOWER CAMBRIAN

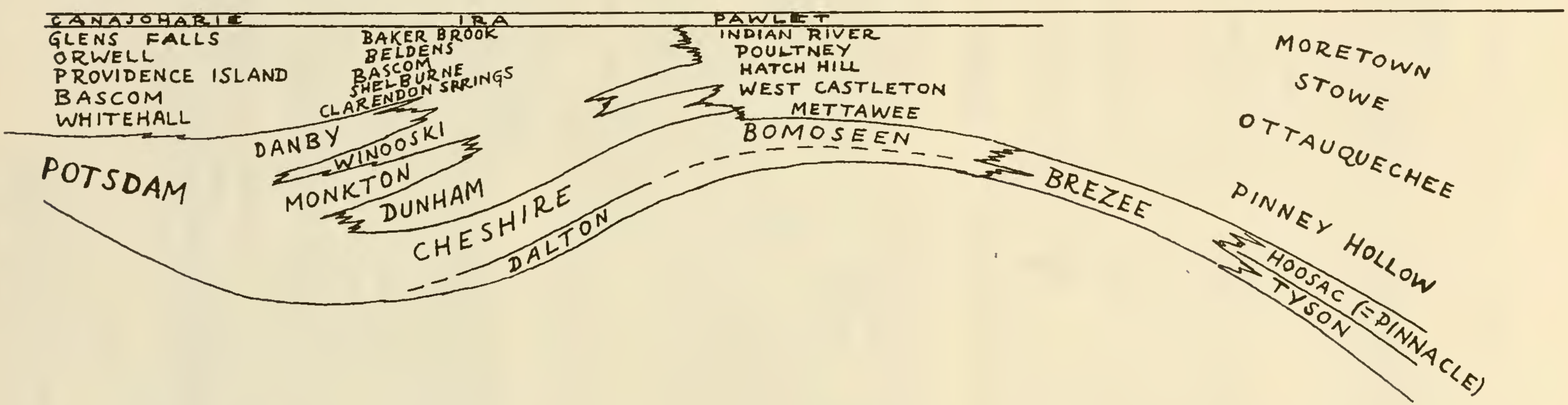
Monkton Quartzite: Beds up to five feet thick of greenish, schistose quartzite, locally containing pebbles of blue quartz several millimeters in diameter, alternating with rusty-weathering dolomitic quartzite, locally cross-bedded and ripple-marked, and beds of light gray dolomite weathering yellow or orange-red. Minor rock types include dark blue-gray dolomite and green or gray-black phyllite. — 300

	Approximate Thickness In Feet
Lower Cambrian fossils have been reported from two localities in the Monkton Quartzite (Kindle and Tasch, 1948; Tasch, 1949; Shaw, 1962).	
Rutland (Dunham) Dolomite: Light gray, yellow-weathering dolomite with thin siliceous partings and minor beds of dark gray dolomite. Upper part (Mallett Member) contains beds up to eight or ten feet thick of sandy dolomite or dolomitic sandstone with conspicuous cross-bedding.	900
A few Lower Cambrian fossils have been reported from the Rutland Dolomite (Cady, 1945, p. 530: Shaw, 1954, p. 1035-1037; Stone and Dennis, 1964, p. 35; Theokritoff, 1968, p. 14).	
Cheshire Quartzite: Massive white or blue-gray vitreous quartzite, locally with cross-bedding, ripple marks, and vertical organismal burrows (Skolithos). Beds of tan-weathering schistose, feldspathic quartzite, and dark gray to black quartzose phyllite are increasingly abundant toward the base. The lower part of the formation corresponds to the upper part of the Mendon Series of Whittle.	1000-1600
Rare Lower Cambrian fossils have been reported from the Cheshire Quartzite (Cady, 1945, p. 528; Shaw, 1954, p. 1034; Stone and Dennis, 1964, p. 28)	
Dalton Formation: Schistose graywacke, conglomerate; minor beds of phyllite or schist, locally with chloritoid, and impure dolomite in the upper part. Approximately equivalent to the Pinnacle Graywacke of Clark, Nickwaket Graywacke of Keith, and the lower part of the Mendon Series of Whittle. The Dalton Formation, although unfossiliferous in Vermont, is considered to be Lower Cambrian because it can be traced into strata on Clarksburg Mountain near North Adams, Massachusetts, from which trilobite fragments have been reported (see discussion in Thompson, 1967, p. 71).	50-300
PRECAMBRIAN	
Mount Holly Complex: Gneiss, schist, amphibolite, quartzite, marble, and calc-silicate granulite, with cross-cutting bodies of pegmatite, gneissic granite, and massive amphibolite. The rocks of the Mount Holly Complex were metamorphosed and deeply eroded before deposition of the Cambrian formations.	unknown

West

East

Schematic west-east section showing inferred relationships of Cambro-Ordovician shelf and basin sequences in west-central Vermont and adjacent New York.



Trip 7: Road log (part 1)

Quadrangles: Chittenden 7 1/2 Minute Proctor 7 1/2 Minute, Rutland 7 1/2 Minute and West Rutland 7 1/2 Minute

Mileage

0.0 Assembly point: On Route 103 in East Clarendon, Vt. at a point 2.3 miles east of intersection of Routes 7 and 103 where the railroad, the Appalachian Trail, and the Long Trail all cross Route 103 (Rutland quadrangle).

STOP 1. Follow trail markers on Appalachian Trail northward to power line (beyond phone line). Turn left (west) and follow power line. The outcrops are in gneisses containing ptygmatically folded, sheared, and crushed pegmatites. Note pale green saussuritization of plagioclase and blue iridescent color of the quartz. These rocks have been referred to the Mount Holly Complex (the town of Mount Holly is a few miles to the northwest).

Other rock types are seen westward along the power line: migmatite and amphibolite.

The unconformable base of the Dalton Formation can be seen at the highest crest on the power line (Note the damage already done by indiscriminate hammering - PLEASE use no hammers here; the weathered surface shows the unconformity better). Note quartz pebble conglomerate overlying a seven foot zone that may be in part a regolith or simply non-conglomeratic Dalton. Development of magnetite in the adjacent part of the Mount Holly is taken as indicative of weathering prior to deposition of the Dalton.

Return to cars and proceed west along Route 103.

2.3 Intersection of Routes 7 and 103: turn right (north) onto Route 7 toward Rutland.

7.1 Intersection of Routes 7 and 4 west (West Street): proceed straight ahead on Route 7 north.

7.2 Intersection of Routes 7 and 4 east (Woodstock Avenue): proceed straight ahead on Route 7 north.

Enter Chittenden quadrangle.

8.3 Intersection of Route 7 with Field Avenue immediately after A&W rootbeer and Wee Tee Golf course: turn left (west) onto Field Avenue and proceed to end of Field Avenue.

Mileage

8.8 End of Field Avenue: turn right (north) onto Grove Street.

9.0 STOP 2. Bridge across East Creek. Pull off pavement immediately after crossing bridge.

Walk north to power line and follow power line to west (left). A gray dolomite breccia with light gray fragments in a dark gray matrix is exposed along the power line. This rock is referred to the Rutland Dolomite.

Just east of the bend in the power line near the northwest corner of the golf course, Cheshire Quartzite can be seen in near contact with the Rutland in the woods between the power line and the golf course.

Return to bridge across East Creek. Walk east approximately 200 yards along footpath on south bank of stream to just below dam. Rutland Dolomite is exposed here and includes a bed containing hyolithids just below the dam.

Return to cars and reverse direction. Proceed south along Grove Street.

Reenter Rutland quadrangle.

10.1 Intersection with Crescent Street (traffic signal): turn right (west) onto Crescent Street.

10.4 Bridge across East Creek.

10.5 Intersection with Fairview Street: turn left (south) onto Fairview Street.

10.9 Cross State Street and bear half-right.

11.2 Join Route 4: proceed straight ahead westward.

Enter West Rutland quadrangle.

12.2 Intersection of Routes 3 and 4 in Center Rutland: turn right (north) onto Route 3 toward Proctor.

12.3 STOP 3. Quartzites probably belonging to the upper part of the Dalton overlain by gray dolomite also assigned to the Dalton and overlain, in turn, by phyllites and quartzites assigned to the Cheshire (quartzites may be seen in the railroad cut above).

The Pine Hill thrust is situated between Stops 3 and 4.

Mileage	
13.0-13.2	STOP 4. Ira Formation cut by a dike.
13.8	Power line
13.9	STOP 5. Outcrops in road and to right are in dolomites and quartzites of the Danby Formation situated on the axis of an anticline.
14.4	STOP 6. Quarry on right in Sutherland Falls limestone of Clarendon Springs Dolomite. Large outcrop to north is in upper dolomite member of Clarendon Springs.
14.6	STOP 7. Riverside quarry in Shelburne Marble on left (west).
14.8	Large outcrop on right of road in upper member of Clarendon Springs Dolomite. Note dike.
14.9	Intersection: turn right (east) onto old road.
15.7	STOP 8. Outcrops east and west of road are of Winooski Dolomite overturned to the west. Crest of ridge east of road is held up by quartzites at the top of the Monkton, also overturned to the west. Outcrops of the Monkton east of the ridge crest are in the axial region of the Cox Mountain anticline which plunges southward at this locality.
15.8	Intersection: turn right onto Route 3 toward Proctor.
16.6	Intersection: turn left onto bridge over Otter Creek.
	Sutherland Falls to right.
16.7	Turn right.
16.9	Vermont Marble Company exhibit. Most of the outcrops in the village of Proctor are in Winooski Dolomite
17.1	STOP 9. Park on right. Walk west along railroad track. The first outcrop on the right is of interbedded quartzite and cross-bedded sandy dolomites belonging to the Danby and the lower member of the Clarendon Springs.
	Quarry just northwest (follow spur track to right) is Sutherland Falls Quarry, type section of the Sutherland Falls member of the Clarendon Springs Dolomite. Dolomites to west of quarry belong to the upper member of the Clarendon Springs. These dolomites are referred to by the quarrymen as the "Intermediate" dolomite

Mileage	
	because they separate the Sutherland Falls Marble from the overlying "Columbian" marble of the Shelburne. The Shelburne crops out in the western part of the village.
	Return to the intersection of Routes 3 and 4 in Center Rutland.
22.0	Intersection of Routes 3 and 4: turn right (west) onto Route 4.
23.7	Junction Route 4 and alternate Route 4: follow alternate Route 4 into West Rutland.
23.9	Intersection alternate Route 4 and Route 133: proceed straight ahead.
24.1	West Rutland High School on left: turn into yard behind school.
	STOP 10. Large outcrop southeast of building is in blue-gray limestone and dolomites referred by Zen (1964) to the Beldens.
	Leave school yard and turn right toward intersection of alternate Route 4 and Route 133.
24.6	Intersection alternate Route 4 and Route 133.
24.8	Junction Route 4 and alternate Route 4: turn right for Route 4 west toward Castleton and Fair Haven.
32.2	Exit.
32.3	Stop sign: turn left to alternate Route 4 and Castleton.
32.7	Stop sign: turn right toward Castleton.
33.3	Castleton: Post Office.
35.0	Castleton Corners: intersection Route 30.
36.4	Hydeville: Lake Bomoseen Inn.
	Enter Thorn Hill quadrangle.
37.9	Intersection: bear left on Route 4.
38.3	Fair Haven village green: intersection Routes 4 and 22A north.
38.7	Intersection Routes 4 and 22A south: begin road log part 2 mileage 0.0.

Trip 7: Road log (part 2) Quadrangles: Thorn Hill N.Y.-Vt. 7-1/2 Minute and Granville N.Y.-Vt. 7-1/2 Minute

Mileage

0.0 Fair Haven, Vermont. Intersection of Routes 4 and 22A at Vermont Structural Slate plant. (Note: there two locations in Fair Haven where routes 4 and 22A intersect; the road log begins at the more southerly of the two, between the Castleton River and the Delaware and Hudson tracks).

STOP 1. Large outcrops on both sides of tracks southeast of highway intersection.

Walk south along 22A to railroad crossing and then left (east) along tracks to first outcrops. These show dark gray graywacke (Bomoseen Graywacke) geometrically overlying purple slate with thin interbeds of quartzite (Mettawee Slate). The contact is gradational.

Proceed westward along Route 4.

1.7 State Line: Bridge over Poultney River (Note: from this point, Route 4 generally heads southwestward).

3.8 Intersection with Fair Haven Turnpike (County Road 21)

4.1 STOP 2. Pull off on wide right shoulder. Caution: this highway carries fairly heavy traffic - don't become a paleontological specimen.

The long roadcut on both sides of Route 4 is mainly in thin gray laminated bedded limestones interbedded with dark gray shale with silty laminae (West Castleton Formation). One such bed of limestone yielded members of the Elliptocephala asaphoides fauna (Theokritoff, 1964, loc. 58-5).

Mettawee Slate is exposed at the northeast end of the cut (Northeastward from about 12 feet southwest of the 'Cattle Crossing' sign). The contact is partly covered but appears to be sharp.

The West Castleton Formation here probably lies in the core of an overturned syncline with an eastward dipping axial plane; on the western limb, the West Castleton grades down into a green argillaceous slate (Mettawee Slate) seen southwest of the white post near the drain on the southeast side of Route 4 opposite the northeast end of guard fence on the northwest side of the highway.

The most southwesterly outcrops in this roadcut (on the southeast side of road) are of dark green silty micaceous subgraywacke which should probably be referred to the Bomoseen Graywacke.

Mileage

4.6 STOP 3. The road cut shows a laminated micaceous silty graywacke (Bomoseen Graywacke). A bed of quartzite is exposed at the northeast end of the cut on the southeast side of the highway.

4.8 STOP 4. Laminated micaceous silty graywacke (Bomoseen Graywacke)

5.1 STOP 5. Outcrop on southeast side of highway: Gray-green argillaceous slate bearing mica spangles; some ankerite nodules (Bomoseen Graywacke).

View ahead of Adirondacks.

5.3 Intersection of Route 4 with Beckwith Road. Turn left (south) onto Beckwith Road.

Skyline half-left is of western margin of Taconic klippe (Bomoseen Graywacke).

6.2 STOP 6. Outcrops in left bank of dark gray contorted argillaceous shale (Snake Hill Shale).

6.3 Intersection of Beckwith Road with Route 273. Turn left (east) onto Route 273.

6.8 STOP 7. Outcrops in hillside on left (south) side of road. Autochthonous carbonate rock geometrically overlain by dark graywacke (Bomoseen Graywacke). The Taconic thrust may be located between the two.

7.1 Intersection.

7.4 Outcrops on both sides of road of Bomoseen Graywacke.

8.1 East Whitehall: Intersection of Route 273 with County Road 21. Turn right (south) toward Granville.

8.4 Outcrops on left side of road of dark gray shales with thin interbedded white-weathering dark gray sublithographic limestone. (A member of the Poultney Slate.)

9.2 STOP 8. Outcrops of dark gray shales with thin interbedded limestone in stream to left (east) of road.

Proceed a few yards south along road to outcrops on right (west) roadside. These outcrops consist of green-weathering dark-gray banded argillites with a few thin dolomitic laminated sandstone layers. (B- member of the Poultney Slate).

The lithologic assemblage seen at Stop 8 is similar to that in the type-section of the Schaghticoke Shale in the Hoosick River.

Mileage	
9.8	Intersection: turn right (west).
10.0	STOP 9. Magenta and bluish-green siliceous argillites, glazed in appearance. These beds are referred to the base of the Indian River Slate and occupy the center of a syncline at this locality. Green and dark gray banded argillites (B- member of the Poultney Slate) occur a few yards to the west and east and their gradation into the Indian River Slate may be observed. Note the light buff weathering characteristic of the Poultney Slate in the outcrops on the north side of the road west of the Indian River outcrop. Note the development of cleavage across the center of the syncline.
10.2	STOP 10. Outcrop on right (north) side of road of dark gray shales with interbedded thin white-weathering dark gray sublithographic limestone (A-member of the Poultney Slate). At the west end of the outcrop, there is an anticline; the shales weather into very thin slabs and are probably very close to the contact with the underlying Hatch Hill Formation.
10.8	Intersection: turn left (south)
11.2-11.3 and 11.7-11.9	Outcrops of Bomoseen Graywacke at left roadside.
12.0	Intersection: turn left (south) Enter Granville quadrangle.
12.4	Intersection: turn left (east). The rise ahead is underlain by Bomoseen Graywacke. The knoll immediately south of the farm marks an outcrop of fossiliferous Lower Cambrian conglomerate (Theokritoff, 1964, loc. 58-12).
13.1	Intersection: bear left
13.3	STOP 11. Walk back to outcrops on bend in road near intersection. Rusty-weathering sooty black shales with rotten-weathering bluish dolomitic sandstones named Hatch Hill Formation (type locality). The shales have yielded <u>Callograptus</u> sp., <u>Dendrograptus</u> sp., and rarely <u>Dictyonema</u> sp. at this locality (Theokritoff, 1964, loc. 55-9). Berry (1961, p. 224) referred this fauna to the Late Cambrian. Proceed easterly along road to outcrop on south (right) side of road of greenish-gray banded argillite with a few thin quartzite interbeds (B-member of the Poultney Slate). This locality has yielded graptolites referred

Mileage

by Berry (1961, p. 225, Ps-4) to the Early Canadian (Late Tremadoc):

- Adelograptus aff. A. divergens (Elles and Wood)
- Adelograptus cf. A. pauxillus (Benson and Keble)
- Adelograptus simplex (Tornquist)
- Adelograptus sp.
- Clonograptus aff. C. timidus Harris and Thomas
- Clonograptus sp.
- Didymograptus sp.

13.4 - 3.5 Outcrops of Poultney Slate on right roadside.

13.6 Outcrops of red and bluish-green argillites (Indian River Slate).

14.2 Note quarry and waste dumps in Indian River Slate to right.

14.2 - 14.4 Section from Indian River Slate through Poultney Slate into Hatch Hill Formation.

14.4 STOP 12. Outcrops of sandstone with interbedded black shales (Hatch Hill Formation). Note easterly dip and cross-bedding indicating inversion at this locality.

14.6 Intersection: turn left (north)

15.3 STOP 13. (Theokritoff, 1964, loc. 55-6). Outcrop on left roadside of fossiliferous limestone-conglomerate. Such limestone conglomerates probably represent masses of limestone boulders that have been carried eastward into the basin from the sand-carbonate shelf (Bird and Theokritoff, 1967).

Limestone boulders at this locality have yielded:

- Serrodiscus speciosus (Ford)
- Fordaspis nana (Ford)
- Fordaspis sp.
- Calodiscus lobatus (Hall)
- olenellid fragments
- Helcionella cf. H. subrugosa (d'Orbigny)
- Stenothecoides cf. S. elongata (Walcott)
- Hyolithes sp.
- Hyolithellus micans (Billings)
- Coleoloides cf. C. prindlei Lochman
- brachiopods

Proceed north. Browns Pond on left. Reenter Thorn Hill quadrangle.

16.4 Intersection: Proceed straight ahead (north) to East Whitehall. Turn left (west) onto Route 273 and proceed to junction with Route 4. Follow Route 4 through Whitehall into Fort Ann. Turn right in Fort Ann (flashing light) onto Route 149. Follow this highway to intersection with Route 9. Turn left (south) and follow signs for Interstate

87 south. Follow Interstate 87 into Albany.

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Trip 8

HYDROGEOLOGY AND REMOTE SENSING OF THE MOHAWK RIVER VALLEY NEAR SCHENECTADY, NEW YORK

by

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Trip Objectives

The objectives of this field trip, not ranked in order of importance, are as follows:

1. To see the deposit in the field which is capable of producing a major ground water supply (in this case 20 to 30 million gallons per day to the City of Schenectady - Town of Rotterdam, New York).

2. To examine the distribution of the unconsolidated deposits associated with the Schenectady-Rotterdam aquifer.

3. To outline geographically the physical boundaries of the aquifer system.

4. To visit the Schenectady pumping facility in order to see the type of facility required to supply ground water to a major population area.

5. To examine the cone of depression associated with the Schenectady-Rotterdam pumpage and the aquifer recharge relationships which result.

6. Demonstrate the effects produced by surface structures (Lock 8-New York State Barge Canal) and the effect exerted on the ground water reservoir by seasonal control of river level.

7. Examine infrared imagery and radiometer data of the field trip area and discuss its hydrologic significance.

Setting

Geomorphic: The field trip area, shown in Figure 1, is a small portion of Eastern Schenectady County. The area lies across the western boundary of a lowland that is bounded on the north by the Adirondack Mountains, on the east by the Taconic Mountains, on the south by the Helderberg Escarpment, and on the west by hills that lie between the Helderberg Escarpment and the Adirondack Mountains. (Winslow et al, 1965). The upland to the west, the lowland to the east and the Mohawk River Valley are the major topographic features in the area This trip will be entirely in the Mohawk Valley flood plain.

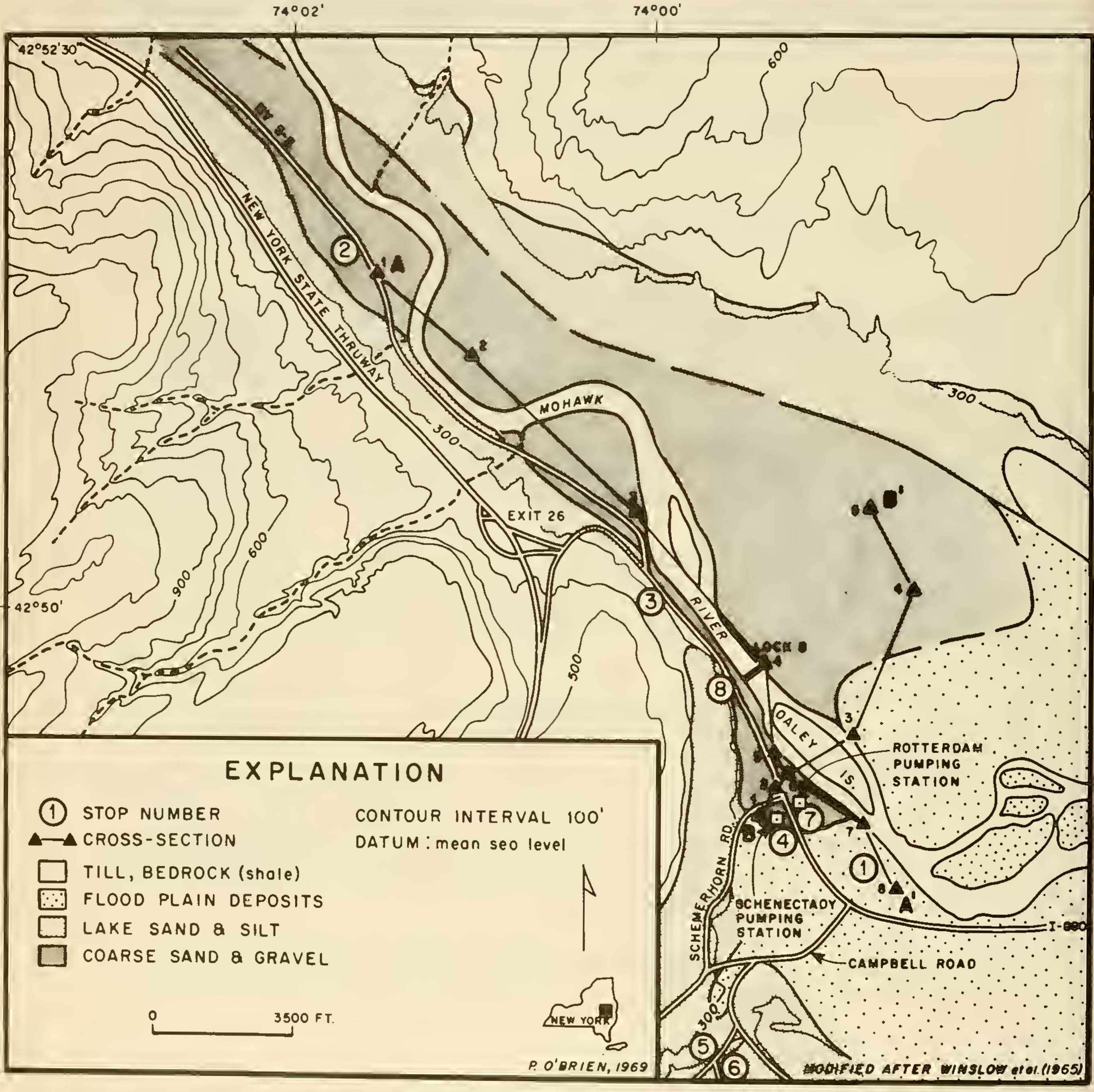


Figure 1. Surficial geology of the Mohawk River Valley west of Schenectady, New York.

Hydrologic: The Capital District area falls within the Hudson River Drainage Basin as defined by Halberg et al, 1964. Schenectady, Scotia, Rotterdam, and Glenville are dependent upon ground water supplies derived from coarse channel gravels located in the Mohawk River Flood plain between Hoffmans and Schenectady, New York. This source is more than adequate to supply the above municipalities combined population of 149,000 (1960 figures), occupying 114.1 square miles. Of the above 149,000 people 109,175 or 73% (population of Schenectady and Rotterdam- 1960) derive their supply from an area about 5 square miles or about 4.4% of the area centered on the Schenectady pumping station (Figure 1).

The Mohawk River flow equals or exceeds 580 million gallons perday 95% of the time. Statements regarding the flow characteristics of the Mohawk are at best generalizations due to the highly regulated flow of the River resulting from hydroelectric power generation at Vischers Ferry downstream from Schenectady, and the maintainance of pool levels for navigation purposes at the numerous Locks in the Barge Canal System.

Stratigraphy and Associated Ground Water Availability

Bedrock

Schenectady Formation (Ordovician): The Schenectady Formation underlies the field trip area and is approximately 2,000 ft. thick. It is composed of grit and sandstone alternating with black and gray argillaceous shale.

The yield of wells in this formation is highly dependent upon the presence of zones of shear and fracture which result in increased permeabilities. Yields are generally 5 gallons per minute or less. Hydrogen sulfide occurrence is common.

Unconsolidated Deposits (Pleistocene and Recent). Please refer to Figures 1 and 2.

Glacial Till: The bedrock hills to the west of the trip area are covered with from 10 to 30 feet of this heterogeneous mixture of rock particles which was deposited directly by the glacial ice. Grain sizes range from clay to boulders

The yield of a well in till will vary from 1 to 5 gpm. A dug well of large diameter, capable of storing a large volume of seepage water in non-use periods, is most successful. Permeable zones frequently occur at the interface between the till and bedrock.

Flood Plain Deposits: These deposits, averaging 20 to 30 feet in thickness are located in the Mohawk River Valley. Their distribution is partially omitted on Figure 1 so that the lake sand and silt and the coarse sand and gravel distribution could be shown. The flood plain deposits are chiefly silt and sand with included organic matter and result from overbank deposition and the reworking of earlier valley fill materials.

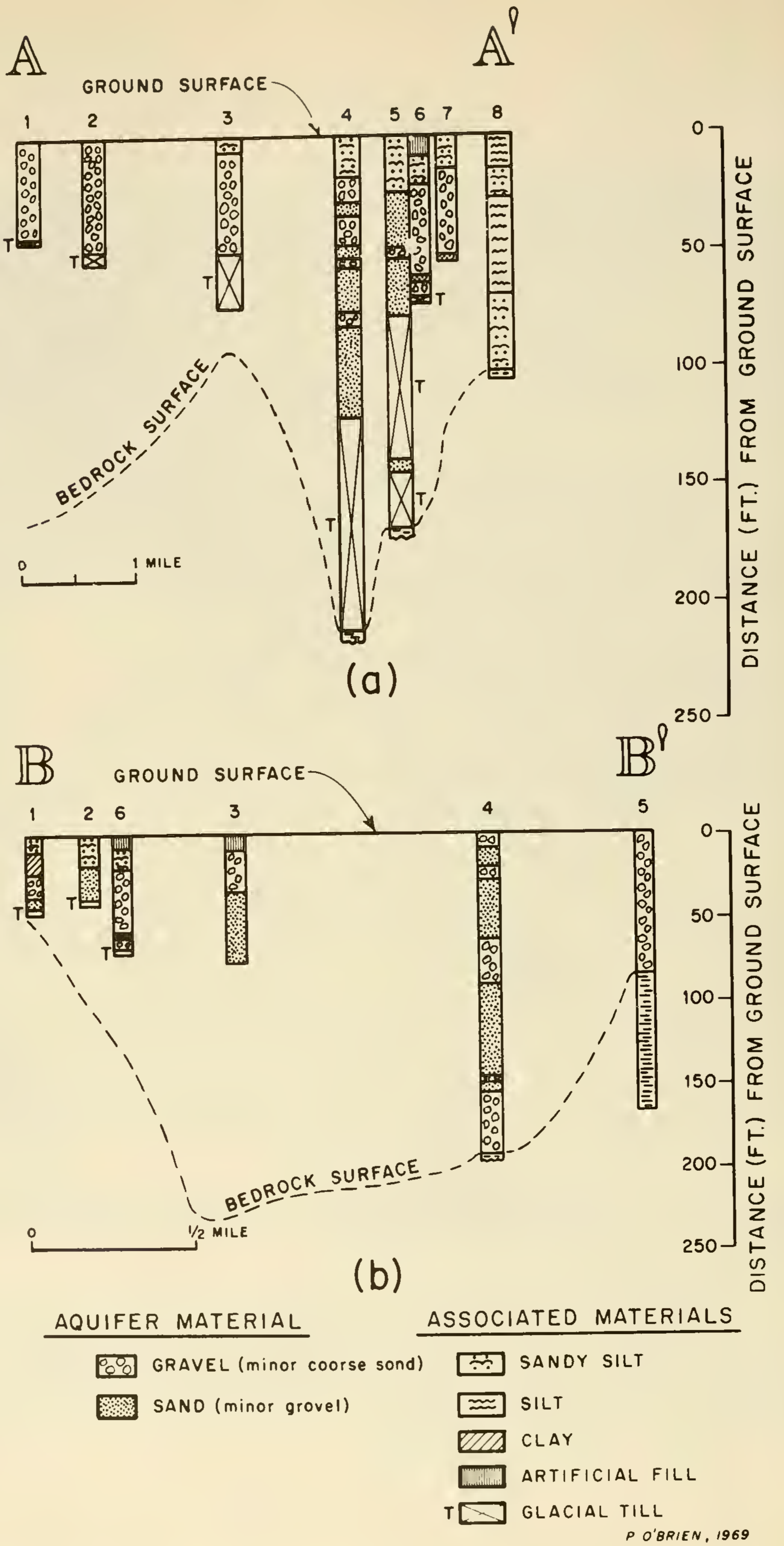


Figure 2. Correlation cross sections, Mohawk River Valley (see Figure 1 for locations).

Yields of such wells range from 20 to 100 gpm. The more prolific wells generally are completed in sand and gravel lenses which are scattered through the deposit. Wells drilled in proximity to a recharge source (i. e. Mohawk River) may, by pumping, induce infiltration from the River which will result in a greater yield.

Lake Sand: These deposits are medium to coarse grained in the higher parts of the deposit and medium to fine grained in the lower part. The sands are predominantly silty. In the Mohawk Valley they were deposited when the Valley served as an inlet to glacial Lake Albany.

Yields range from 5 to 20 gpm and fill domestic and small municipal needs.

Coarse Sand and Gravel: These deposits are by far the most significant source of ground water in the Schenectady area and in the Capital District as a whole. The deposit consists of alternations of well sorted gravel (1/2 ft. to 4 ft. thick) with beds of sand and gravel (up to 70% sand). The degree of sorting varies producing gravel beds with little sand and conversly. The thickness of these deposits varies from 50 to 70 feet with a maximum of about 100 feet.

Yields of wells completed in these deposits are as great as 3500 gpm. This is possible due to the high permeability of the deposits and to the effect of induced infiltration from the Mohawk River caused by the pumpage at the Schenectady and Rotterdam facilities. The impact of induced infiltration will be discussed below.

Hydraulic Characteristics of the Bedrock and Unconsolidated Deposits

Table 1, which follows, summarizes two important hydraulic characteristics, permeability and transmissibility, of the deposits described above and shown in Figures 1 and 2. Permeability is defined in two ways, one as the "coefficient of permeability" (P) and two as the "field coefficient of permeability" (P_f). Both are defined as the rate of flow of water in gallons a day through a cross section of 1 square foot under a hydraulic gradient of 100 percent; however, the former is defined at 60°F and the latter at the prevailing temperature of the ground water. The coefficient of transmissibility (T) is the rate of flow of water in gallons per day through a section of aquifer 1 foot wide and having a height equal to the saturated thickness of the aquifer, under a hydraulic gradient of 100 percent and at a temperature of 60°F. The coefficient of transmissibility is equal to the coefficient of permeability times the saturated thickness of the aquifer.

TABLE 1

Material		P or P_f*	T**
		gpd/ft.^{2}	gpd/ft.
Bedrock			
Schenectady Fm.	P	0.4	50
Unconsolidated Deposits			
Glacial Till	Pf	0.004-0.001	--
Flood Plain Deposits	Pf	5,000	100,000
Lake Sand			
Upper Part	P	20-100	--
Lower Part	P	1000	--
Coarse Sand & Gravel			
Rotterdam Jct.	Pf	100,000	5,000,000
Schnectady Chem. Co.	Pf	15,000	600,000
Town of Glenville	Pf	35,000	1,200,000
City of Schenectady (North of Lock 8)	Pf	1,600	55,000
City of Schenectady (At Well Field)		--	5,180,000
Scotia Naval Depot	Pf	400	60,000
Scotia Village	Pf	20,000	800,000

* Coefficient of permeability and field coefficient of permeability, respectively, as defined above.

** Coefficient of transmissibility

Induced Infiltration

General: The cone of depression created by pumping adjusts itself in order to supply the water demanded by pumping. If aquifer permeability is low the cone will enlarge in order to intercept a volume of water adequate to satisfy pumpage. To satisfy the same pumpage a smaller cone will develop if the aquifer is of high permeability, all other factors equal. If the cone intersects a recharge boundary, such as the Mohawk River, it will cause water to flow towards the pumping site thus inducing infiltration.

Schenectady-Rotterdam Aquifer Boundary Conditions: Boundary conditions are seldom ideal and "all other factors" are never equal and hence a cone of depression symmetrical about the pumping center is not observed in nature. The cone may intercept a "no flow" boundary, a recharge boundary, or a boundary due to changing permeability in the aquifer materials.

The bedrock valley wall on the west side of the field trip area is essentially a no flow boundary and is approximately marked by the 300 ft. contour of Figure 1. Some side hill seepage in the colluvium probably does reach the aquifer through the partially confining beds above the aquifer but the contribution is negligible compared to recharge from the Mohawk River.

The depth of the Mohawk channel is sufficient to penetrate the gravel aquifer and hence is a recharge boundary. Evidence for the hydraulic connection between river and aquifer can be obtained by comparing hydrograph records at the Schenectady pumping facility with river level data. The levels do rise and fall in nearly direct phase with each other. Further, peak river discharge due to floods also are reflected in aquifer levels. In addition, one finds, from measurement of aquifer temperatures in the large number of observation wells available in the Schenectady-Rotterdam pumping area, a nosing of temperature contours pointing from the river towards the pumping wells. This reflects river water infiltration. Winslow et al (1965) made monthly maps over a period of one year showing that the nosing temperature pattern remained fairly constant but varied in absolute amount as the river temperature changed from about 32°F to 74°F in the climactic year.

The change in temperature mentioned above has a considerable effect on recharge to the aquifer in that water at 32°F requires a hydraulic gradient about double that at 80°F to transmit the same amount of water. Hence, the inverse relation of viscosity and temperature is significant in ground water flow.

To the south of the Schenectady-Rotterdam well field area the aquifer rapidly changes facie to a silty sand composition as shown by drillers logs and becomes less permeable as demonstrated by "stacking" of water table contours.

Relatively impervious till and/or bedrock underlie the aquifer and only partially confining silts and clays overlie it. Hence we are dealing with a water table aquifer.

Infiltration Sites:

General: The location of infiltration is governed by the aquifer boundary conditions previously described. By previous argument the principal source of recharge is the river. Several factors determine where in the river infiltration will occur. They are:

1. Whether it is navigation or non-navigation season of the year (the former from mid-April to mid-December and the latter the balance of the year).

2. The relative thickness of river bottom deposits.

3. Hydraulic gradient between river and aquifer.

4. Hydraulic dredging to maintain navigability.

Upper Pool - Lock 8: During navigation season the upper pool (west pool) at Lock 8 is about 14 feet deeper than the lower pool. In the course of a year the bank interval in the upper pool between high and low yearly water levels is exposed and scoured by river action. During navigation the 14 ft. pool creates a considerable hydraulic gradient causing the river water to infiltrate to the aquifer which is near the surface at Lock 8. Water levels in wells just above and below Lock 8 show that levels downstream are above river level and hence part of the ground water goes back to the river. Some water also migrates under the Lock structure as underflow. The remainder of the ground water is intercepted by the cone of depression and flows toward the well field.

Lower Pool - Lock 8: During non-navigation season the upper pool level is dropped and there is only 1 ft. of difference in upper and lower pool levels. Upper pool infiltration essentially ceases and the principal infiltration site becomes the area of the lower pool. Two factors control this location. First, during navigation water is constantly pouring over the Lock gates thus keeping fine materials from settling in the lower pool. Second, the underflow is coming to the surface just east of Lock 8 in the lower pool. Hence this area of river bottom is very effectively winnowed resulting in higher permeability. The hydraulic gradient is reversed in non-navigation season as evidenced by well hydrographs, and recharge occurs in the lower pool area.

Downstream from Lock 8: The river bed silts and clay increase in thickness downstream from Lock 8 thus reducing the infiltration to the aquifer from the river due to lowered permeability. However, the river channels are periodically dredged hydraulically to maintain navigable depths. In 1956 dredging operations were conducted in the channel adjacent to the west side of Daley's Island (Figure 1). Thirty-five thousand cubic yards of river bed material were removed. Permeability thus was immediately increased; however, since a deeper channel is created by dredging, siltation probably occurs very rapidly again and hence infiltration capacity in this area once improved by dredging may be rapidly reduced.

Effects of Seasonal Control of River Level on Cone of Depression

Accepting the premise that the principal infiltration sites are above and below Lock 8 as described above, one can predict the behavior of the cone of depression from the following relation:

$Q = PIA$ where, disregarding units,

Q = discharge
P = permeability
I = hydraulic gradient
A = cross-sectional area through which flow takes place

If discharge is to remain the same with the aquifer permeability constant, then the area of the cone of depression must increase if hydraulic gradient decreases. Hydraulic gradient does decrease in the non-navigation season. Further, we know that recharge to the aquifer is significantly reduced during the winter season and hence the cone must expand to meet the requirement of a constant discharge. There is evidence that the cone may extend upstream 1,000 feet from Lock 8 in the non-navigation season. In fact, discharge does not remain constant but decreases on the average through the winter months. Nevertheless, demand is still enough and recharge sufficiently reduced to require an expanded cone of depression.

Maximum Yield of the Aquifer

Maximum yield is the maximum pumpage rate which does not lower the water level below the top of the screened interval in wells drilled to the bottom of the aquifer units. The factors discussed above result in the lowest maximum yield of the aquifer in winter and the highest maximum yield in summer. Summarizing, Winslow et al (1965) have calculated the following maximum yields.

TABLE 2

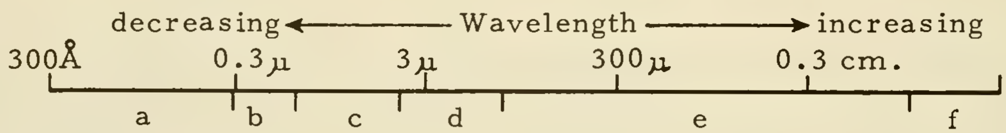
	Millions of gallons per day
Maximum Yield During Navigation Season	60-100
Maximum Yield- Non-Navigation Season	
With Limited Replacement of Storage*	60
With No Replacement of Storage*	30

* 100 day non-navigation season assumed

Winslow et al (1965)

Infrared Imagery and Photography

General: Every physical object on earth which is at a temperature above absolute zero (-459°F) radiates infrared radiation in the 0.72 to 1,000 micron region of the electro-magnetic spectrum. This radiative phenomenon is due to oscillation of the atoms and molecules of natural substances and most significantly the emitted radiation is proportional to the temperature of the emitter. The characteristics of infrared energy vary with wavelength as shown below.



a = ultraviolet
b = visible light
c = near infrared (up to 1 μ)
d = intermediate infrared (3-5 μ)
e = far infrared (8-14 μ)
f = radar

Figure 3 - Portion of the electro-magnetic spectrum including infrared wavelengths.

Near infrared can be photographed whereas intermediate and far infrared is thermal energy to which regular emulsions will not respond. Therefore infrared photography is usable in daytime only due to its dependence on reflected solar infrared radiation whereas infrared imagery is recording emitted radiation and can be used day or night.

According to the Kirchhoff Law, emission of radiation can occur only at wavelengths where absorption occurs. Therefore the infrared transmission capabilities of the atmosphere are important in that they control what wavelengths penetrate the atmosphere. Further, it is known that ozone, carbon dioxide and water vapor in the atmosphere are responsible for creating major infrared absorption bands. Figure 4 shows the transmission spectra of the atmosphere and the "atmospheric windows" which allow selected infrared penetration.

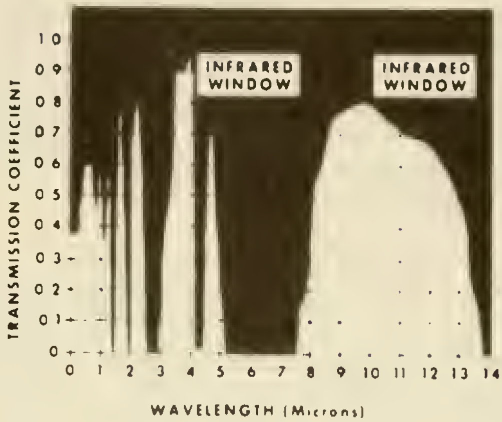


Figure 4. Transmission spectra of the atmosphere.

Schenectady Area Imagery

Infrared imagery of the field trip area was flown at about 7:20 p.m. on August 24, 1968, accompanied by an airborne radiometric survey. This imagery and radiometry was obtained from the 8 to 14 micron or far infrared end of the spectrum. The dashed line in the approximate center of the imagery (Figure 5) is the line along which radiometric readings were made continuously and recorded as a strip chart (Figure 6). The radiometer is a rigid mounted, fixed field device which is boresighted to the center of the line scan. Simultaneous marking of both the imagery and the radiometer strip chart, at one second intervals, allows easy correlation of the two.

Campbell Road and N. Y. State 5-S image warm as does Schemerhorn Road. However, the first two are of greater period and amplitude. Campbell Road and Rt. 5-S are both better exposed to solar radiation than is Schemerhorn Road. The latter is partially covered by tree canopies and is particularly so at the point of the overflight. Campbell Road and Rt. 5-S are both wider than Schemerhorn Road and hence have more exposure to the radiometer. The flight line crossed Rt. 5-S at a very acute angle therefore its period of exposure to the instrument was greater than either of the other roads.

The Mohawk River images warm relative to the land, exclusive of the roads, because water has a greater capacity to retain thermal energy than does land. During the day the river would image cool because water can absorb a larger portion of incident solar radiation than can land.

The imagery shown in Figure 5 is only a small portion of that obtained. On the field trip the whole flight line, which includes the Figure 5 portion, will be displayed. Figure 5 appears to increase in darkness (coolness) from south (left) to north. However, when the radiometer strip chart (Figure 6) is examined it appears that the sections labeled A and C average about the same temperature and thus the gradually increasing south to north darkness of the imagery is due to instrument drift of the imagery system.

The salient feature of Figure 6 is the apparent lower average temperature of section B. Significantly, this area is approximately coincident with the Schenectady-Rotterdam aquifer. It is not possible, at this time, to say that the coolness of section B is is due to the aquifer. Since the groundwater at this time of year is much cooler than the surface and if that effect is propagating upwards one would expect the superjacent soils to be cooler. However, this is not proof that the imagery is revealing the aquifers presence. Additional work is currently underway which hopefully will clarify this point.

See pages 8-14 and 8-15 for Figure 5 and Figure 6

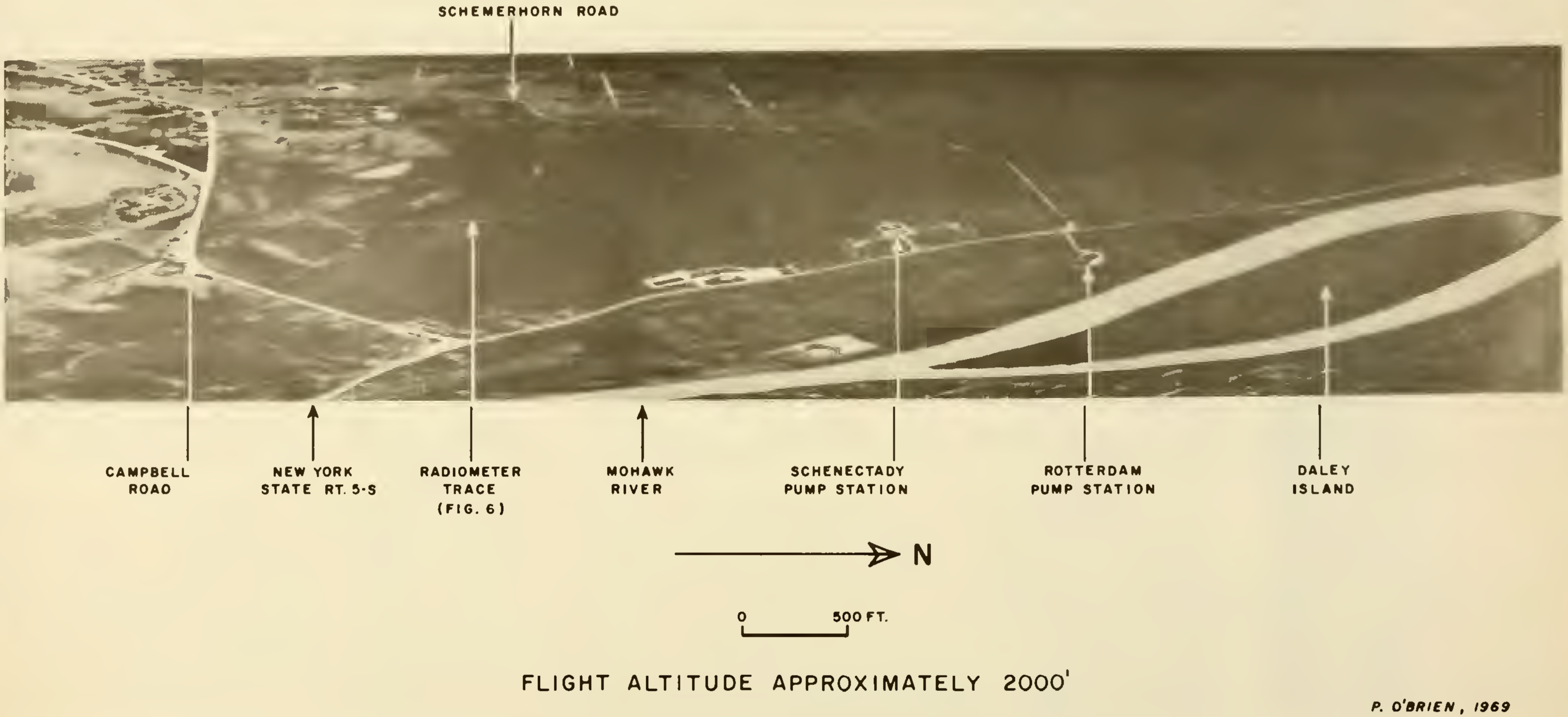


Figure 5. Far infrared imagery from flight August 24, 1968 near

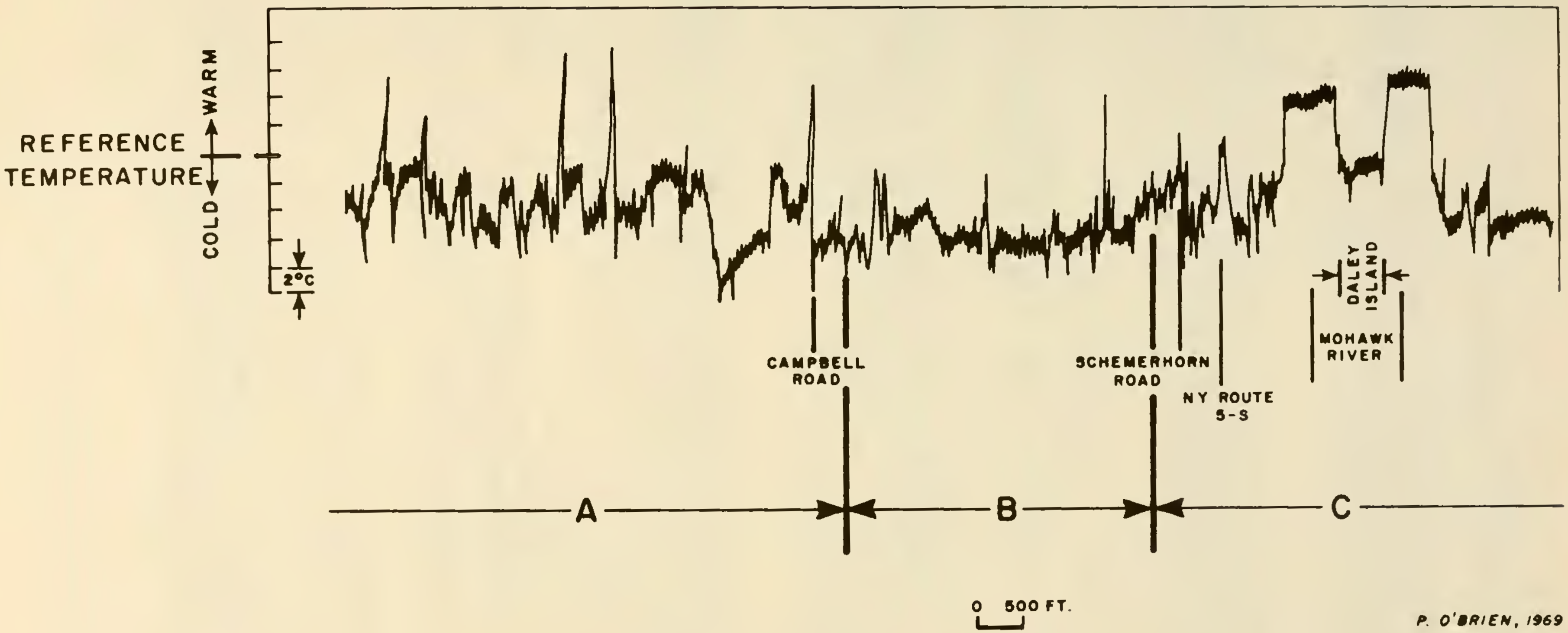


Figure 6. Airborne radiometric temperature profile from flight August 24, 1968 near Schenectady, New York. Courtesy HRB-Singer Inc.

Trip No. 8: Mileage Log

Leave Thruway Motor Inn, Washington St., Albany, entering the New York State Thruway* at Interchange No. 24. Travel north. Leave Thruway at Schenectady Interchange No. 25. Mileage begins here.

<u>Mileage</u>

00.0 Interchange No. 25, New York State Thruway. Follow I-890 traveling over the flat topography of the sand plain.

2.5 Schenectady City Line.

2.5 Begin descent to Mohawk River Flood Plain.

4.5 Panoramic view to west of till covered bedrock hills which border the flood plain.

5.2 Enter area covered by Figure 1. Mohawk River on right; General Electric turbine-generator building on the left.

5.9 I-890 terminates. Feeds directly into N.Y. State 5-S.

6.0 Turn right into dirt parking area.

6.4 <u>STOP 1</u>
Follow one-lane dirt road from parking area towards river. Pit from which top soil has been removed to expose aquifer materials. The gravel disappears a short distance south of this location.

0-8"	grass mat
8"-6'4"	clay and sand layers with latter predominant (sand lenses to 7")
6'4"-12'	sand and medium gravel
12' +	floor of excavation; top of coarse cobble zone; top of Schenectady-Rotterdam aquifer.

9.9 Return to highway; continue west on N.Y. 5-S, passing abandoned Erie Canal structure on the right.

10.8 Entering township of Lower Rotterdam Junction.

11.3 <u>STOP 2</u>
Left on gravel road just before crossing Boston and Maine RR tracks. DeLuke Sand and Gravel Pit. The working, high-wall in this pit shows a considerable thickness of the sandy-gravel facies. A hydrologic problem at this pit which resulted in court litigation will be discussed from a hydrologic point of view.

*No stopping is allowed on the New York State Thruway.

13.7 Return to highway; East on N.Y. 5-S past Interchange 26.

14.0 STOP 3
Pull off highway into a small parking area to the right. Large exposure of Schenectady shale. Joint sets are prominent.

14.4 Go east on N.Y. 5-S past Lock 8, Erie Barge Canal.

15.1 STOP 4
Schenectady Pumping Station. Examination of the pumping facilities; Discussion of the hydrologic implications of proposed highway construction in the vicinity of the station. Review and discussion of the infrared imagery will be conducted at this stop.

15.5 Follow N.Y. 5-S east to Campbell Road. Turn right on Campbell Road.

16.0 Fork in road; bear left.

16.4 Pass beneath Delaware & Hudson RR bridge.

16.5 STOP 5
Turn right on Burdeck St., Stop at Land-Fill site. Brief discussion of the land-fill operation and related hydrologic problems. Could this operation in any way jeopardize the Schenectady-Rotterdam supply?

16.6 West on Burdeck St. to fork in road; bear left on Thompson St.

16.8 STOP 6
Pull off, park on berm to the right. Cross highway for overview of large pit operation which has resulted in a serious erosion problem.

18.1 Return to N.Y. 5-S via Thompson, Burdeck and Cambell Rd. Turn left on N.Y. 5-S.

18.7 STOP 7
Pull off in a small parking area on the left side of highway at Schemerhorn Road. Cross highway. Brief look at Rotterdam Pumping facility. Opportunity to take a closer look at an old lock structure of the abandoned Erie Canal, known in the early 1800's, when it was constructed, as "Clinton's Ditch".

19.2 STOP 8
Continue west on N.Y. 5-S to Lock 8, Erie Barge Canal; pull off in parking area to right. Discussion of infiltration phenomenon due to presence of the lock structure.

19.9 West on N.Y. 5-S to Interchange No. 26; enter New York State Thruway southbound towards Albany.

20.9 Outlet channel of Early Lake Amsterdam thru which the Thruway now passes.

21.8 Outcrop of interbedded Schenectady shale and graywacke.

21.9 Leave area covered by Figure 1.

22.2 West shore of glacial Lake Albany.

Continue south on N.Y. State Thruway to Albany Interchange No. 24. Thruway Motor Inn via Washington Avenue. Trip terminates. Total trip mileage about 45 miles.

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CARBONATE FACIES OF THE HELDERBERG GROUP (LOWER DEVONIAN) OF NEW YORK

Leo F. Laporte
Brown University

General remarks. Inasmuch as each participant has received a recent reprint discussing the stratigraphy, paleontology, and inferred depositional environments of the Helderberg (Laporte, 1969) it is unnecessary to repeat all that here. Instead I will provide you with an itinerary of the four localities we will be visiting as well as some of the major features or problems of Helderberg paleoecology that will be discussed at each stop.

Publications that serve as background for this trip are Rickard (1962) and Laporte (1967). Indeed, were it not for Dr. Lawrence Rickard's excellent critical re-study of Helderberg stratigraphy my work and that of my students would not have been possible.

Itinerary. Trip begins at intersection of Routes 9W and 396, some seven miles south of Albany. Go west 2.6 miles along 396 and turn south (left) along South St. for about 0.4 mile; turn into Callanan quarry.

> STOP 1 - Loc. 53a*, Delmar 7 1/2' quad.
> Outcrop on south wall of quarry of upper Silurian Rondout Fm. and lower Devonian Manlius Fm. lying unconformably on middle Ordovician Normanskill Fm. greywackes. (Higher Helderberg units exposed elsewhere in quarry.) Section records late Silurian-early Devonian transgression which occurred throughout central Appalachians. Here, both the Rondout and Manlius have a tidal flat aspect: laminated, mudcracked, dolomitic limestones and dolomite; some horizons, particularly in the Manlius have abundant scour surfaces with limestone-pebble conglomerates. Algal stromatolitic structures common within the Manlius - both irregular laminations sometimes swelling into small hemispherical masses as well as free-lying algal oncolites (concentrically algal laminated clasts or skeletal debris). The Rondout has few if any fossils while the Manlius has abundant tentaculitids, leperditid ostracodes, ramose ectoproct bryozoans, scattered snails, and U-shaped burrows. Note rapid vertical changes in carbonate rock types and difficulty of tracing individual horizons any significant distance. What we have here, in short, is a fossil shoreline.

Return north (left out of quarry) along South St. to 396 and turn west (left). Continue on for about 1.1 mile to four corners. Turn north (right) toward Feura Bush; stop at roadside exposure about one mile north of four corners intersection. Despite pastoral nature of this stop beware of traffic.

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*Localities numbered and located in Rickard, 1962.

STOP 2 - Loc. 53b, Delmar 7 1/2' quad.
Outcrop along road of Manlius Fm. with massive stromatoporoid bed interfingering with fossiliferous pelletal muds. Beds below (around outcrop in woods) are thinly laminated, mudcracked and dolomitic. The undulating contact between the stromatoporoid bed and the pelletal mudstone suggests truncation by erosion of the underlying bed before the stromatoporoid unit was deposited (although the contact may be somewhat modified by compaction). I have interpreted this as a tidal channel formed in subtidal pelletal muds in which stromatoporoids flourished. This is analogous to modern channels in tidal flats where abundant oysters or mussels grow. Lateral migration of the channels causes the stromatoporoid beds here, and elsewhere in the Manlius, to appear as thin, laterally continuous time-stratigraphic horizons. Intimately associated with these subtidal horizons of the Manlius are the dolomitic, mudcracked, poorly fossiliferous laminated beds which indicate periods of subaerial exposure, algal mats, and occasional flooding by unusual events (storms, monsoonal tides, or perhaps spring lunar tides). Thus, the Manlius displays a "complex facies mosaic" of tidal flat and shallow subtidal horizons, caused by deposition near mean sea level. Slight fluctuations in sediment accumulation caused significant shifts in evironmental regimen. Migration of flats, tidal channels, and shallow subtidal areas results in the present day complex internal stratigraphy of the Manlius.

Turn around and return to four corners. Turn west (right) and continue on 396 for about 7 miles to the village of Clarksville where 396 joins 43 at "T" intersection. Turn northeast (right) and stop at roadside cut, 1 1/2 miles beyond village.

STOP 3 - Loc. 56, Clarksville 7 1/2' quad.
Outcrop of upper Manlius at low end of outcrop, up through the full Coeymans (about 30 ft.). The Coeymans is a pelmatozoan calcarenite (or biosparudite or grainstone) with tabulate corals, various ectoproct bryozoans, and brachiopods, expecially the pentamerid Gypidula coeymanensis. The Coeymans records deposition in a shallow, well-agitated, subtidal environment, seaward of the Manlius tidal flat-lagoon. Anderson (1967) has recognized four distinct facies within the Coeymans: Bioturbite facies, Cross-stratified facies (both seen here), Sheet-deposit facies and Bioherm facies (seen elsewhere, somewhat higher in the Coeymans). These are interpreted by Anderson as recording integrading environments from a somewhat deeper shelf to a shallower shelf interspersed with mud shoals and intershoal channels or with patch reefs.

Turn around and return 12.4 miles to 9W via 43 and 396. Turn south (right) on 9W and continue for about 15 miles. At route 81 turn west (right) and go about 1.3 mile. Outcrop on left of upper Coeymans, Kalkberg, and most of New Scotland formations.

<u>STOP 4</u> - Loc. 48, Coxsackie 15' quad.
Coeymans exposed at east end of outcrop. Grades up into Kalkberg (boundary placed at first bed of continuous chert) which is a carbonate mudstone (biomicrite) with a great abundance and diversity of marine shelly invertebrates. Brachiopods, ectoprocts, and trilobites are especially common. Pelmatozoan debris is also present, being more abundant in the lower part of the unit. Thin shaly interbeds and chert layers are common. The unit is highly burrowed. The New Scotland Fm. lies transitionally over the Kalkberg and is more argillacaeous and quartz silty. The fauna of the New Scotland is somewhat less abundant than the Kalkberg although perhaps more diverse; like the Kalkberg the New Scotland is also highly burrow-mottled. Toward the west end of the roadcut the New Scotland is deformed (small folds and thrusts). The Kalkberg and New Scotland record open shelf environments seaward of the Coeymans. The New Scotland had significant amounts of fine-grained terrigenous detritus entering from the northeast and is more properly considered an argillaceous, quartz silty mudstone alghough there are some lime-rock interbeds.

These four stops show the vertical sequence of Helderberg environments: 1) the tidal flat and restricted shallow subtidal (Manlius); 2) open, higher energy, subtidal (Coeymans); 3) open, lower energy, subtidal carbonate shelf (Kalkberg); and 4) the open, lower energy, subtidal clastic shelf (New Scotland). These environments coexisted from west to east, for several tens of miles; following the regional transgression, they migrated westward across New York State forming the stratigraphic sequence we have seen today.

GEOLOGY OF THE SOUTHERNMOST ADIRONDACKS

by

James McLelland
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INTRODUCTION

The area under consideration is shown in figure 1. It contains most of the Gloversville and Lassellsville 15' quadrangles. In addition it includes the southern halves of the Piseco and Lake Pleasant 15' quadrangles. Until recently, much of this area has received very little geological attention. Miller (1921-22) reconnaissanced the Gloversville quadrangle, but his map was never published. Miller's 1916 map of the Lake Pleasant quadrangle provides little pertinent information. The same may be said of his 1909 map of the adjoining Broadalbin quadrangle (which the present author is currently mapping).

In 1937 Cannon published his well known study of the Piseco Dome. However, by his own admission, he did little more than a very broad reconnaissance of the gneisses comprising the southern half of the quadrangle.

More recently there has been mapping activity to the west, north, and east of the area under discussion. Nelson (1968) has examined the Ohio 15' quadrangle, and Thompson (1955-59) studied the Harrisburg quadrangle. Bartholomé (1956) mapped in the northern half of the Lake Pleasant 15' quadrangle. In 1959 and 1960 Isachsen ran some reconnaissance traverses through the Lassellsville 15' quadrangle.

At present Duane Wohlford of Oneonta State College is mapping in the Lassellsville 15' quadrangle. The area mapped by Wohlford overlaps with that mapped by the author. This situation has arisen because neither man was aware of the other's presence until June of 1969. Dr. Wohlford has kindly provided me with information

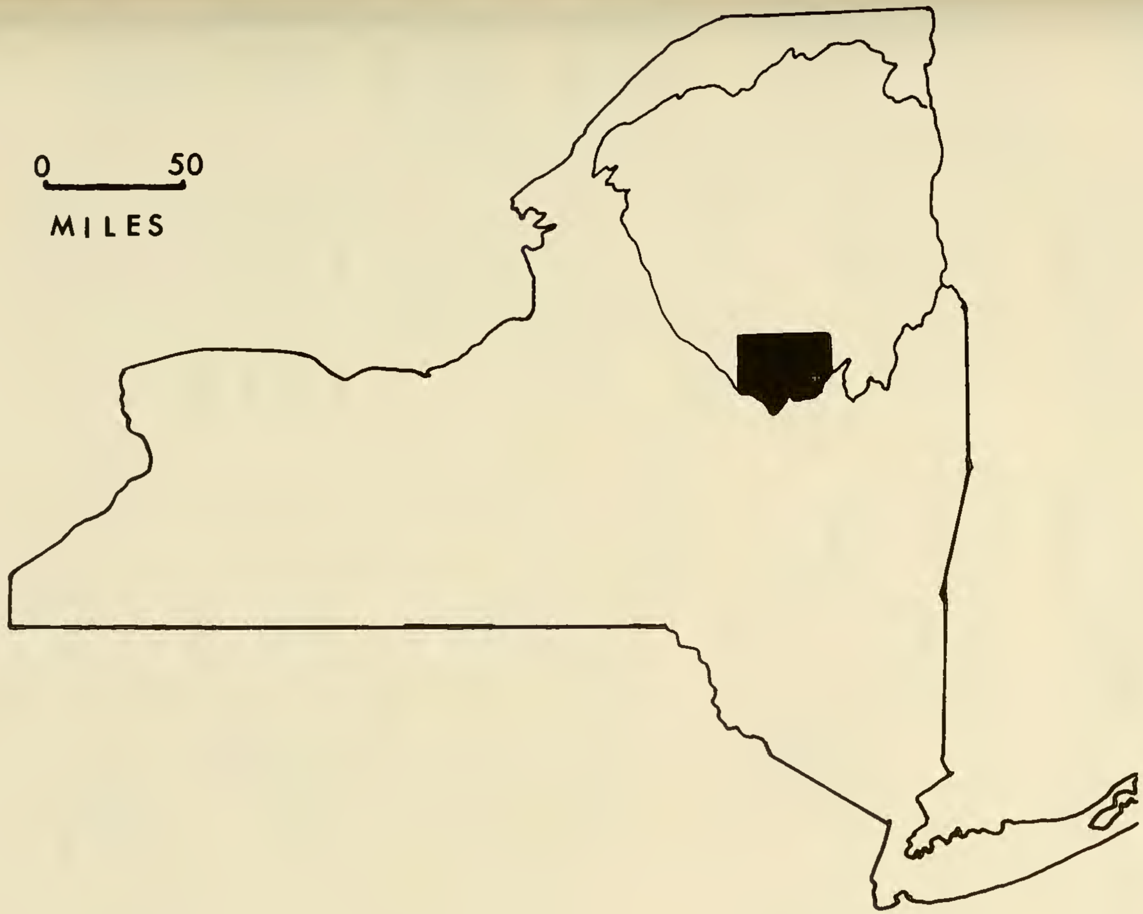


Fig. 1. Blackened area shows approximate location of Precambrian units discussed here.

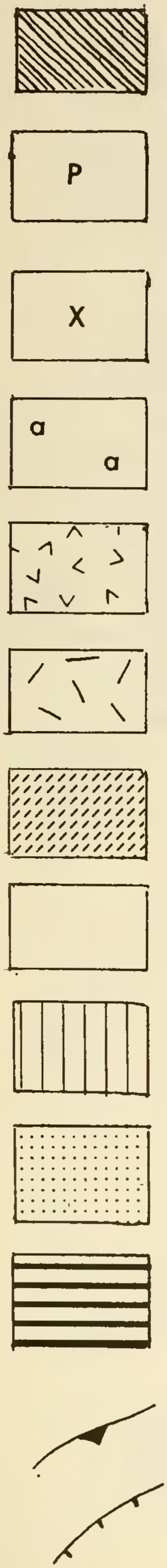
concerning the contacts of the thick quartz monzonite gneiss unit where it crosses the westernmost part of the map area. This information has been incorporated in figure 2.

Until rather recently, geological understanding of the Adirondacks has been severely hampered by certain conceptual biases. The first of these is that the vast majority of quartzo-feldspathic rocks are igneous in origin. In the Adirondacks this belief led early investigators to suppose that the region had been intruded by vast quantities of syenitic and/or granitic magma (Miller, 1923). Adherence to this model led investigators to produce maps with a style now referred to as "blob" geology. From the present author's own experience, "blob" geology can easily result if one fails to recognize stratigraphy and to follow contacts. Such myopia is easily explained in those who believe that the Adirondacks have been "cut to pieces" by a great batholith of quartzo-feldspathic magma. A corollary appears to be that strike and dip information is superfluous, since foliationsin "xenoliths" of gneiss merely reflect the shape of the underlying batholith. Alternatively, a "xenolith" may be rotated, etc., and; therefore, its foliation represents an irregular detail. As a result of such axioms, there exist a number of Adirondack quadrangle maps that are without a single strike and dip.

A second and more recent conceptual bias dictates that, aside from basic rocks, significant intrusives are absent from the Adirondacks. This school of thought is characterized by the belief that, until proven otherwise, all gneisses are paragneisses or meta-volcanics. Actually this is an attractive view and one which the author used to adhere to. Unfortunately strong adherence to this belief can lead to very serious mapping error. This is because quartzo-feldspathic meta-intrusives do exist in the Adirondacks, and failure to recognize them as such makes it very difficult to interpret certain contacts or pinchouts. The resulting interpretation may be marked by some very exotic fold structures which probably do not exist.

The matter of conceptual bias has been stressed, because such bias is impossible to avoid. In reality, whenever we map

LEGEND



Glacial Cover

Paleozoics
Undifferentiated

Olivine Metagabbro
Grades into amphibolite and/or pyroxenite

Amphibolite and Pyroxenite

Hornblende-Biotite Quartz Monzonite Gneiss
Orthopyroxene present locally. Both equi-and inequi-granular.

Quartz Diorite Gneiss

Migmatite
Includes bands of quartz-biotite-oligoclase gneiss

Garnetiferous Quartz-Biotite-Oligoclase Gneiss
Includes minor quartzites and leucogneisses

Leucogneiss
Includes minor quartzite and pelitic units

Charnockite and related
Quartz-Mesoperthite Gneiss

Quartzites and feldspathic quartzites
Includes minor pelitic units

Dip of foliation drawn at contact

Normal faults. Hachures on down thrown side

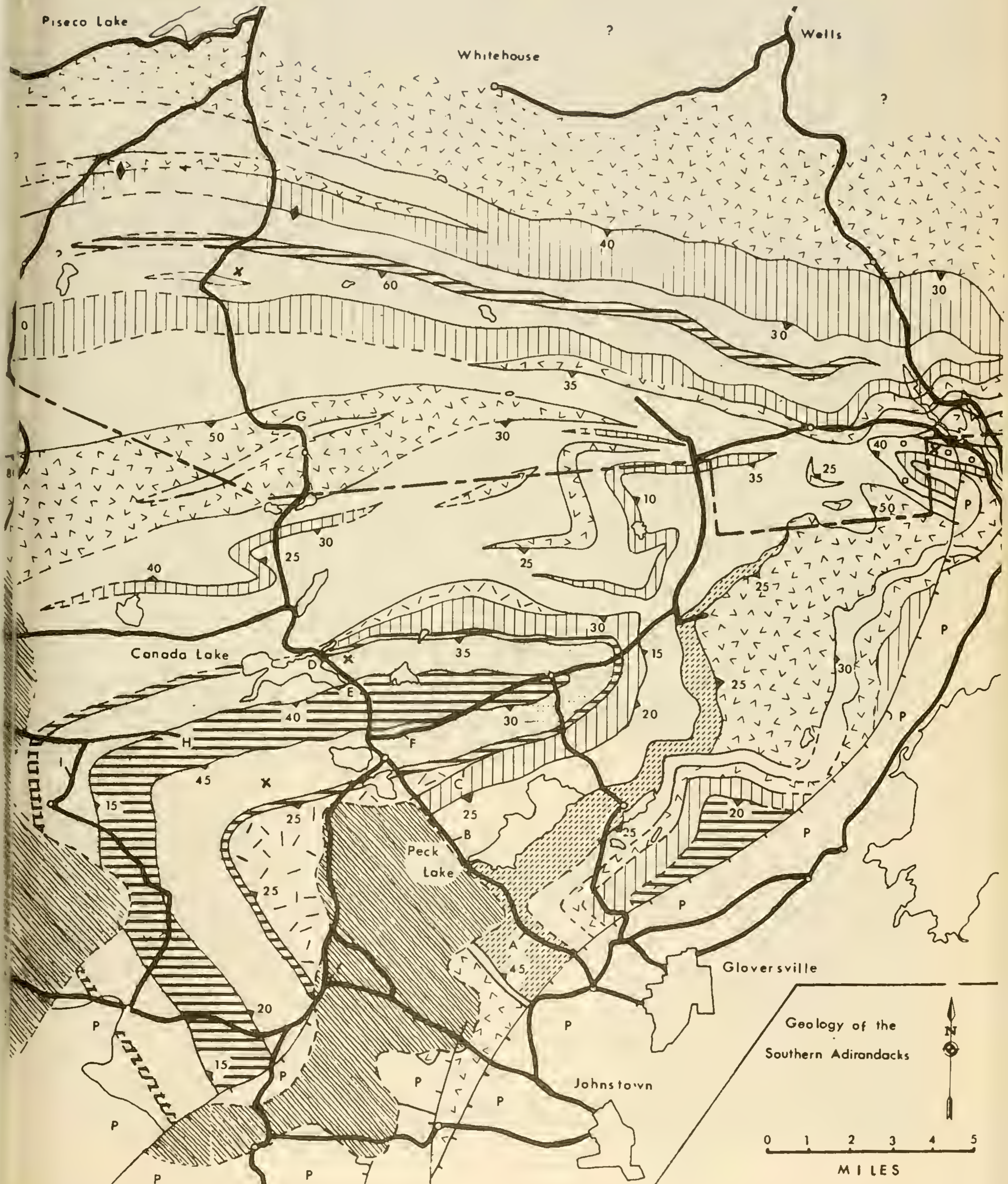


Fig. 2. Geologic map of the Southern Adirondacks

we are guided by hypotheses. Inevitably the final results are influenced by our individual conceptual framework -i.e.- our imaginations. To deny this influence is to delude ourselves. A much more fertile approach is to admit the existence of biases and to evaluate those operating upon us. The result can be a more accurate, and realistic, geologic map prepared under conditions which are enjoyable. This has been the approach used in the work described herein. In particular, the author has accepted the possible existence of ortho - and paragneisses of all kinds. Also posited was the existence of a traceable stratigraphy on which is superimposed several fold events -at least one of which is isoclinal and recumbent.

Lithology and Stratigraphy

Within the area, and in adjoining quadrangles, there exists a distinct and mappable stratigraphy. In fact, the stratigraphic divisions currently recognized are believed to represent the minimum number of mappable units. Continued research will eventuate in greater refinement. In addition, it must be pointed out that the scale of figure 2 does not permit some contacts to be shown.

In the following text headings are by lithology and individual bands (formations?) are discussed beneath the relevant heading.

Quartzite

Without any doubt, quartzites are the most important rocks in the area. Where they are sufficiently thick and continuous, they serve as excellent marker beds. Such is the case with the three major quartzites shown on the map in figure 2. Strictly speaking, the term "quartzite" has a very narrow lithological definition. In the southern Adirondacks the strict definition of quartzite applies only to relatively pure 6 inch to 20 foot thick bands of white or rose quartzite. These are intercalated with quartzo-feldspathic, pelitic, or basic bands. In some instances pure quartzite is subordinate to feldspathic and even garnetiferous quartzite. None the less, the units shown as quartzite in figure 2 are all characterized by the fact that about 50%, or more, of their lithology is comprised of legitimate quartzite beds.

The three major quartzite units are described below. Approximate thicknesses are given in parentheses.

Irving Pond Quartzite (~2000')

This is by far the largest quartzite unit in the area. It consists of interlayered quartzites and quartzo-feldspathic gneisses. Basic bands are subordinate except in local situations. Much of quartzite lithology is comprised of alternating quartzite and darker pelitic bands; the latter being about 1/8"--1/4" thick and giving the rock a distinctive appearance reminiscent of rhythmic layering.

Garnet is pervasive throughout the unit and exhibits the pale lavender color that is distinctive of garnets associated with quartzites in the southern Adirondacks. Thin sections show that the garnetiferous bands contain some 5% - 10% sillimanite.

Very near its outer contacts the Irving Pond quartzite becomes pelitic and biotite rich. This change affects some 50 feet of exceedingly variable lithology. Some of these pelitic gneisses are well exposed in the roadcuts at the east end of Canada Lake (see stop E). Within this outer lithology, garnets tend to become smaller and dark wine-red in color.

The Notch Quartzite (~200')

This northernmost quartzite band is very similar to the Green Lake Quartzite. Stratigraphically, it can be followed over a distance of at least 15 miles. A portion of this unit outcrops along Route 10 between Pine and Piseco Lakes and near the side road to Shaker Place.

Woodworth Pond Quartzite (~100')

This band is exposed on the southeastern border of the area. It is similar to the Green Lake and Notch quartzites.

Green Lake Quartzite (~100')

This unit contains a large quantity of feldspathic material, but there is still a sufficient quantity of quartzite to justify mapping it as such.

Typically, the unit consists of 1" - 6" alternating bands of quartzite and feldspathic quartzite. The weathered surface exhibits a distinctive white color tinged with a very pale green. Lavender garnets abound.

Locally this band is highly sheared and fine grained. Megascopically these bands appear black and glossy. Microscopically, they are quite definitely mylonitic. This is the case at stop D.

The most distinctive thing about the Green Lake Quartzite is its stratigraphic position and continuity. Coupled with its resistance to erosion, it provides an excellent marker formation.

The Green Lake Quartzite is most highly sheared just east of Canada Lake. Here the shearing and mylonitization appears to be related to the intrusion of a nearby olivine metagabbro. Elsewhere shearing is not as intense.

The Green Lake Quartzite is an excellent marker and greatly facilitates distinguishing the Royal Mountain Quartz Diorite from the Hogback Mountain Charnockite.

Charnockite

As in the case of the quartzites, not everything mapped as charnockite is charnockite in the strict sense. Most of these rocks are granitic and quartz-syenitic gneisses that exhibit only a minor development of orthopyroxene. The most prominent mafic is green hornblende. However small amounts of orthopyroxene are present in most thin sections.

The most consistent mineralogy of the charnockite sequence is the ubiquity of mesoperthite. Although orthoclase, microcline, plagioclase, or antiperthite occur locally, mesoperthite is present to some extent in almost every thin section. Often the mesoperthite is visible in a stained hand specimen viewed with a 10 X hand lens.

There are two recognizable charnockitic bands in the area. These are described below.

Hogback Mountain Charnockite (~ 3000')

This rock is magnificently exposed in the large roadcuts at the east end of Canada Lake (stop D).

On the relatively fresh surface of the roadcut the unit weathers dark green, to light green, to pink. On the weathered surface, it tends to be a light reddish brown. Specimens broken open with a hammer exhibit a light brown coloration. Black or bronzish hornblende are seen. Bronzish orthopyroxene is also present and is difficult to distinguish from hornblende in hand specimen.

Both mega - and microscopically, quartz grains are seen to be extremely elongated. This is a typical feature of the unit.

The Hogback Mountain charnockite possesses a rather well developed foliation but lacks much pronounced compositional banding. Many pink "bands" turn out to be discontinuous and fairly irregular color "swaths". Locally, amphibolite layers cause banding to develop.

A very distinctive pink microcline aplite is associated with the Hogback Mountain charnockite. It occurs in several 10'-20' thick bands; one of which is well exposed just south of Canada Lake.

An approximate mode of the charnockite is given in Table A. Note that garnets are not present in the charnockite.

Cathead Mountain Charnockite (~2000')

This unit differs from the Hogback Mountain charnockite in that the latter is by far the more mafic of the two. For the most part the Cathead Mountain charnockite is a leucogneiss containing less than 5% mafics. Where mafics do occur, they are often chloritized. However orthopyroxene does appear locally.

The feldspar of this unit is generally mesoperthite, and this is consistent with the mineralogy of the Hogback Mountain charnockite.

The Cathead Mountain charnockite is well exposed in the roadcuts along Route 30 some 1.8 miles north of where the road crosses the Sacandaga River near the town of Northville. It is equally well exposed in the brand new roadcuts along Route 10 near the side road to Shaker Place between Pine and Piseco Lakes. The on-strike distance between these two points is approximately 17 miles. The unit has been traced for a total on-strike distance of 23 miles and can be projected another 14 miles through the Ohio 15' quadrangle.

An approximate mode is given in Table A.

Leucogneiss (~2500' total)

Layers of leucogneiss are widespread throughout the area. They consist of white to rusty weathering, well banded garnetiferous quartz-feldspar gneisses. For the most part these rocks contain sillimanite at approximately 5% by volume. The bulk of the leucogneisses consist of quartz, oligoclase, and K-feldspar. Quartz is generally present in quantities of 40-50% by volume. Sodic oligoclase and K-feldspar each account for about 25% by volume. Garnets are usually small, well-rounded, and light red to lavender in color. Except for local variations, biotite is present in amounts not exceeding 10% by volume. Microscopic examination shows the biotite to be red to red-brown in color. In outcrop a large number of granitic layers are found banded together with the lithology described above. Quartzite layers are common but thin.

Leucocratic gneisses may be found in almost any section of gneisses richer in biotite. However, not all of these leucocratic bands are continuous, and most seem to merge transitionally into biotite rich pelitic gneisses. This feature probably represents sedimentary facies changes in the original rocks. Most bands of this type have not been mapped, but four are large enough to be shown in figure 2. Two of these outline what appear to be drag folds north of the F_1 fold core. A third occurs in the Northville syncline at the eastern edge of the map. The fourth is a small body near the center of the map.

By far the largest body of leucocratic gneiss envelopes the Cathead Mountain charnockite. It is typically white in color and highly garnitiferous. Quartzite layers are common. This unit is well exposed on the Old State Road 2 miles north of where Route 30 crosses the Sacandaga River near Northville. The total thickness of this unit appears to be in the order of 2000 - 3000'.

In the central map area the southern contact of this unit is graditional into garnetiferous biotite-quartz-feldspar gneisses.

Garnetiferous Biotite-Quartz-Oligoclase Gneiss (~7500')

This is the most widespread paragneiss in the area. It occurs in an extensive band north of Canada Lake. Reconnaissance traverses indicate that this same unit is widely exposed in the Broadalbin 15' quadrangle and that these rocks swing around so as to fit conformably into the F_1 fold pattern.

The biotite-quartz-oligoclase gneisses are distinguished by the grey color of their weathered surface and by their black, biotite rich interiors (up to 20% biotite). Scattered about in this matrix are variable quantities of white or pink K-feldspar augen.

Unlike the leucogneisses, these darker gneisses do not usually contain sillimanite. Exceptions are known, especially in the roadcuts along Route 29A between Pine Lake and Stratford. In such instances associated biotite is red and both K-feldspar and deep wine-red garnets are common. This relationship suggests a genetic relationship between the phases involved. Reactions such as those described by Lundgren (1962) or Wones and Eugester (1965) are of relevance.

In outcrop these gneisses exhibit variable banding. Intercalated units may be leucocratic paragneisses or granitic layers. Some of the latter may be anatectic. There are many instances where banding is much subdued or almost absent.

There are several features that complicate the subdivision of this unit. In the first place, the lithology is overwhelmingly monotonous,and mappable discontinuities are not easily located. Secondly, most of the recognizable discontinuities die out over short distances. This is the case with many of the leucocratic layers within the quartz-biotite oligoclase gneiss. In addition to these matters, the unit has behaved in an incompetent fashion during at least two folding events. As a result, what little stratigraphy exists is grossly complicated by folded flowage folds. Locally there exists a rodding which obliterates foliation. Such difficulties are compounded by the fact that these gneisses are the most easily intruded and eroded in the area. The net result is a very difficult and frustrating problem in stratigraphy and structure. This problem has not been satisfactorily resolved, and, as a result, it is not certain whether or not the biotite-quartz-oligoclase gneisses are repeated by an unrecognized major fold. If such a fold exists, its axis runs about N70W between Canada Lake and the large quartz monzonite band to the north. Such a fold would greatly alter thicknesses represented in the stratigraphy.

In general the garnetiferous quartz-biotite-oligoclase gneiss appears to be quite similar to Engel and Engel's " least altered" gneiss. An approximate mode is given in Table A.

Migmatite (~0 - 500')

A single mappable unit of migmatite is found in the eastern and southern part of the area. It is possible to argue that this migmatite should be mapped as part of the biotite-quartz-oligoclase gneiss to which it is closely related. This argument has merit, but its distinctive stratigraphic and structural properties make the migmatite easily recognizable and clearly mappable.

Lithologically, the migmatite is composed of granitic and biotite granitic bands interlayered with biotite-quartz-oligoclase gneiss. Garnets are not widespread in the granitic layers, but they do occur.

The most striking feature of the migmatite is the remarkable degree to which it has developed minor folds of all generations. In no other lithology is folding so ubiquitous and impressive.

The folding and granitic composition of the migmatite suggest that its origin is in part igneous -either due to anatexis or the effects of the nearby quartz monzonite intrusive.

Amphibolite and Pyroxenite (~ 0 - 300')

Small bands of amphibolite and pyroxenite occur in all the rock units described hereunto. None of these bands are large enough to map.

Within the Northville Syncline, there exists a large, mappable unit of mixed amphibolite and pyroxenites. These occur as in thick basic bands as well as interlayered with quartzo-feldspathic metasediments. The origin of these basic rocks in uncertain, but they may be related to the large olivine metagabbro that occurs on the north limb of the syncline (see description under Metaigneous rocks).

Metaigneous Rocks

Royal Mountain Quartz Metadiorite (~0 - 2500')

This distinctive, medium grained, unit consists of andesine plagioclase (An_{40}), quartz, and a little green hornblende. The quartz content averages around 30-40%. Generally the rock is well lineated but banding is poorly developed. Where banding does exist, it is usually the result of 2"-3" layers of amphibolite in the metadiorite. In places these amphibolite layers are broken, twisted, and disrupted giving a typically igneous appearance. Such cross cutting relationships are well developed in the large ledge just across from the Canada Lake Store and Post Office (stop D).

A distinctive feature of the metadiorite is its white weathering surface. This leads to an extremely leucocratic appearance which can cause confusion with granitic gneisses. On a fresh surface the rock bears a remarkable resemblance to the Hogback Mt. Charnockite. In many instances field identification has required staining by sodium colbaltinitrite. It is probable that a significant percentage of Adirondack quartz syenites are actually misidentified quartz metadiorites.

Rooster Hill Inequigranular Quartz Monzonite Gneiss* (~ 0 - 20,000')

This unit occupies a tremendous volume in the southern Adirondacks. In the past it has been mapped as inequigranular hornblende granitic gneiss (phgs); or inequigranular pyroxene-hornblende quartz-syenitic gneiss (phqs). In fact all of these lithologies represent local variations of the unit here described. In the opinion of the present investigator, it is not generally possible to subdivide the Rooster Hill gneiss into mappable units of hbg, phgs, or phqs. Attempts to do so lead to regrettable losses of time and the expenditure of much frustrated effort.

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* After writing this section, the author read Buddington's (1939) description of the Hermon Granite. There can be little doubt that the Rooster Hill gneiss is closely related to the Hermon Granite.

The most distinctive feature of the Rooster Hill gneiss is the presence of large (1"-4") megacrysts of potassium feldspar. These megacrysts are generally well aligned and define an excellent lineation. In many places shearing strain has led to crushing and flattening of the megacrysts in the plane of foliation. Perhaps most interestingly of all are those places where minimal deformation provides strong visual evidence suggesting that the megacrysts are metamorphic relicts of original phenocrysts. Such exposures are amply available on the large ledges that rise above West Stony Creek southeast of the town of Benson. Here the megacrysts are developed to 4" in length and Carlsbad twinning is clearly seen on their white weathered surfaces. The crystals are clearly euhedral and in many places exhibit a random orientation. In other places crystals line up to form flowage type structures. The appearance is very much that of an igneous porphyry.

The megacrysts in this gneiss include microcline, orthoclase, and microperthite. Microcline is best developed in the more highly sheared varieties. Feldspar color appears to have little correlation with which phase is present. Myrmekite is commonly developed next to the megacrysts.

It has been traditional for Adirondack geologists to refer to this unit as a phase of the syenite or quartz-syenite series. Indeed the rock itself has often been classified as a quartz-syenite (see, for instance, Nelson, 1968, p. F18). This practice is wholly unwarranted and is dangerously misleading. Mineralogically, the rock classifies as a quartz-monzonite gneiss, and should be referred to as such (see Table A). This point is stressed not because the author is a zealous advocate of rock classifications, but because much of the charnockitic series has a quartz-syenitic composition and ought not to be confused with the present gneiss from which it differs genetically.

It is interesting that the quartz-monzonite gneisses have not been found in cross-cutting relationships anywhere within the area. To the contrary a number of conformable contacts are exposed. In addition, there are in the southeastern part of the area many relatively narrow (25'-200') bands of quartz monzonite gneiss which are conformably folded with the paragneisses over distances of several miles. Evidence of this sort appears to be consistent with a non-intrusive origin for the quartz monzonite gneiss. However the shape of some of the large bodies, and relationships to the paragneisses argue against this conclusion. Consider, for example, the large body of inequigranular quartz monzonite gneiss that sits near the area's eastern boundary. At its southwestern extremety, it quite clearly prys the paragneisses apart and thus makes room for itself. Moreover, bodies of paragneiss within the quartz monzonite gneiss are generally cross-cut by pegmatite and granitic veins. These paragneiss inclusions are never continuous but clearly seem to gradually fade away into "wisps".

From the foregoing it is concluded that the quartz monzonite gneiss was synkinematically intruded and that conformity is to be expected. Once again we find that what seems "reasonable" depends mainly on the premises which serve as starting points for reason.

A particularly interesting feature associated with the quartz monzonite gneiss is the metasomatism that appears to have taken place in the vicinity of its contact. Locally the biotite-quartz-oligoclase gneiss exhibits excellent microcline "dents-du-cheval". These are particularly well developed in the gneisses just west of Woodworth Lake. This metasomatism appears to be the result of igneous activity and not the cause of igneous "looking" rocks.

An approximate mode is given in Table A.

<u>Equigranular Hornblende Quartz Monzonite Gneiss</u> (~ 0'-500')

This unit is similar to the Rooster Hill inequigranular gneiss except that K-feldspar megacrysts are absent, and the rock shows distinctive crystals of black hornblende up to $\frac{1}{4}$" across.

Equigranular hornblende quartz-monitic gneiss occurs in narrow bands or near the margins of some of the megacryptic bodies. It also occurs as large, integral bodies in the Piseco Dome and its extension to the southeast.

In some places it appears that the equigranular hornblende quartz-monzonite is nothing more than a highly sheared phase of the inequigranular phase.

Examples of equigranular hornblende granitic gneiss may be seen where the Northville-Lake Placid Trail crosses the West Branch of the Sacandaga River at Whitehouse (Lake Pleasant Quadrangle).

Olivine - Metagabbro (~0'-500')

Within the area there exist four bodies of olivine metagabbro. The best developed of these is located in the northern limb of the Northville syncline. This body is 500' wide at one point and has a lateral extension of well over a mile. Grain size varies from medium to coarse, and relict ophitic texture can be found near the center of the intrusion.

Of particular interest in the metagabbro are the presence of pyroxene-spinel symplectites which permit approximate limits to be placed on the P-T conditions of the imprinted metamorphic event. These symplectites consist of symplectite intergrowths of orthopyroxene, clinopyroxene, and green spinel all which separate embayed crystals of olivine from embayed labradorite. The total assemblage appears to represent Kushiro and Yoder's (1966) reaction :

$$\text{Anorthite} + \text{Forsterite} \rightleftharpoons \text{Orthopyroxene} + \text{Clinopyroxene} + \text{Spinel}$$

By utilizing an extension of curve (B) in their figure 2, we may set the upper conditions of metamorphism at approximately 700-900°C and pressures corresponding to 16-20 km depth. The pressure temperature range can also be narrowed by employing the constraint that P-T conditions must lie on the sillimanite side of the sillimanite-kyanite equilibrium. Dr. Philip Whitney of R.P.I. is currently studying these, and other Adirondack symplectites and coronites, in order to better ascertain the P-T conditions in different parts of the Adirondacks. Dr. Whitney's work indicates that the depth of metamorphism increases northward from the Northville area (personal communication, 1969).

TABLE A

Approximate Modes

	1	2	3	4	5	6	7
Quartz	33	35	25	30	30	42	40
Orthoclase	17				10		
Microcline			28			20	
Microperthite				25			
Mesoperthite	15	50					
Plagioclase	An_{20} 12	An_{20} 11	An_{25} 32	An_{20} 25	An_{20} 30	— 18	An_{40} 52
Hornblende	7		8	10			5
Biotite	1	3	4	15	15	5	2
Sillimanite					3	5	
Orthopyroxene	10						
Garnet					7	8	
Accessories	5	1	3	5	5	2	1

1. Hogback Mt. Charnockite
2. Cathead Mt. Charnockite
3. Inequigranular Quartz Monzonite
4. Inequigranular Quartz Monzonite
5. Garnetiferous Biotite-Quartz-Oligoclase Gneiss
6. Leucogneiss
7. Royal Mountain Quartz-Diorite Gneiss

Structural Geology

Much of the Southern Adirondacks are characterized by low dips. These average near 30°, but dips of 10°-15° are common, and horizontal layering is locally prevalent. Over a rather extensive tract the structure resembles that of a homocline and was interpreted as such by Miller (1923). This homoclinal aspect is only apparent, and is precluded by tight minor folds that occur throughout the region.

As shown in figure 2, southern Adirondack structure is dominated by two major fold sets. The earliest recognizable system is magnificiently represented by the large F_1 fold that sweeps through Canada Lake. The F_2 system is developed at several places in the area, but the most notable axis trends N60W to E-W and passes from the west-central edge of the map to just north of Gloversville. Both sets of fold axes plunge gently eastward at 10°-15°.

The F_1 and F_2 folds exhibit recognizably different styles. The F_1 folds are typically elongate and tightly appressed isoclinal folds. They are nearly always found in reclined or recumbent attitudes. The F_2 folds tend to be more open and often display curvatures akin to a slightly flattened letter S (i.e.- S).

Figures 4 a); b); c); and d); show stereograms of foliations and lineations which exhibit the F_1 and F_2 symmetries. Stereograms a) and b) represent data collected in the general vicinity of Canada Lake. Stereograms c) and d) are for data collected in the general area defined by the F_1 fold nose.

F_1 Fold System

The geometry of the F_1 fold is depicted in figure 3. As indicated in the caption, the dips in figure 4 are somewhat exagerrated in order to emphasize the topology. Thus one may clearly see that while the axial trace of the F_1 fold usually parallels contacts (except at the fold-nose), the axis of ths fold trends northwest. The fold axis trends approximately N20W - N30W in the vicinity of

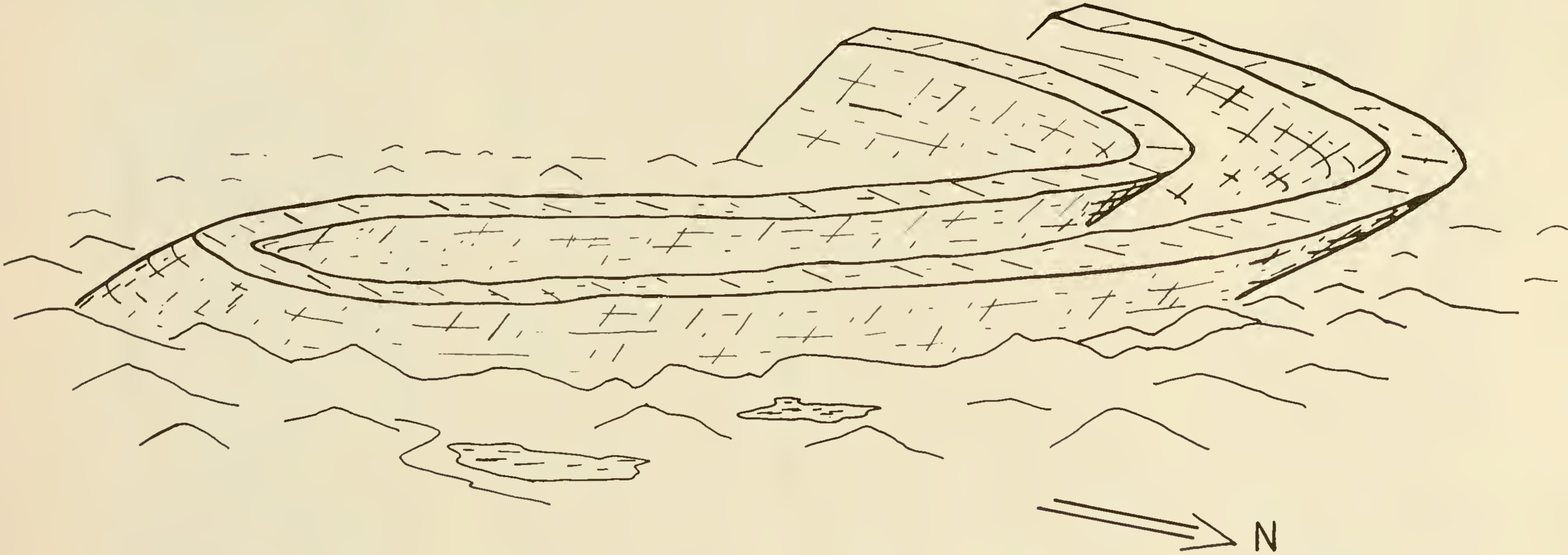

Fig. 3. Schematic block-skeleton diagram depicting fold structure in south-central part of map. The fold is drawn on the layer of Hogback Charnockite. Dips are exaggerated in order to emphasize fold geometry.

the fold nose. Of course, this trend may vary locally depending on the influence of later fold events.

F_1 minor folds are recognizable in many parts of the area. They are especially pronounced in the biotite-quartz-oligoclase gneiss and in the migmatite. The axial trends of a large number of these folds have been incorporated into the lineation stereograms of figure 3. The most impressive F_1 minor folds are found in the Irving Pond quartzites. These are especially impressive where pyroxenite and amphibolite alyers form the fold core; however, care must be taken to distinguish these features from boudinage. As seen here, F_1 folds are remarkedly elongate and show minor crinkles near their noses. Folds and crinkles have trends of N20W - N30W and plunge 10^o-20^o SE. Thus these minor folds replicate the majo. fold; and, in fact, proved to be of great assistance in delineating the major F_1 fold.

Mineral lineation associated with the F_1 system is difficult to find. Most of the mineral lineation shows a consistent trend of N60W to E-W and is clearly associated with the F_2 fold system. However, a few F_1 mineral lineations have been found in the area eastward from Canada Lake to the fold nose. These trend N-S to N30W and plunge gently to the south and southeast. F_1 lineations are generally very weak and could be easily missed among the very dominant F_2 linear elements. Microfabrics may show more instances of relict F_1 lineation.

According to all available information, the large F_1 fold appears to be a reclined or recumbent antiform. Its dimensions qualify it as a major structure in the Adirondacks. In the map area alone it can be followed for 20 miles. Reconnaissance surveys indicate that the same structure extends into the neighboring Broadalbin 15' quadrangle. This relationship suggests that the F_1 fold shown in figure 2 lies at the core of a larger nappe-type structure that incorporates all of the southern Adirondacks.

There exists a problem concerning the position of the F_1 axis in the region where it appears to abut against the large eastern block of quartz-monzonite gneiss. It is not certain if the axis swings northward around the body; or passes through the body. The latter hypothesis is suggested by the approximate F_1 symmetry exhibited by the map pattern of the units of quartz monzonite gneiss. However, this pattern may represent nothing more than the expected result of conformable, synkinematic intrusion. If such is the case, the quartz monzonites would post-date F_1 fold events. This possibility is suggested by the absence of F_1 symmetries in stereograms of quartz monzonite macrofabrics. However, the high degree of homogeneity in the quartz monzonite gneiss makes it very difficult to ascertain their macrofabrics. Therefore, F_1 structures may be obliterated. Hopefully microfabric studies will help to clarify this aspect of the problem.

Upward warping of the F_1 axis is favored by the appearance of tight recumbent minor folding in the paragneisses at the east edge of the quartz monzonite block. However these relationships are not incompatible with a pre-F_1 age for the quartz monzonite; nor are they incompatible with shouldering by intrusion.

It is possible that the F_1 fold axis swings northward and eastward around the quartz monzonite. Such an interpretation gains support from stratigraphic and structural relationships in the area. Strikes in the biotite-quartz-oligoclase gneiss turn north and the unit begins to thin down as it approaches the Northville Syncline. In the vicinity of the syncline the foliation strikes NW and dips to the north. This suggests that a fold nose exists in the area. However, these features can be just as easily interpreted as local convulutions in proximity to the quartz monzonite.

Further support for the northward swing of the fold axis is given by the S shaped folds that are crossed by the county line near the center of the map area. These folds trend about N50W-N60W and are consistent with a westward shouldering due to the intrusion, or fold interference, by the quartz monzonite.

Clearly the matter of the eastward extension of the F_1 fold

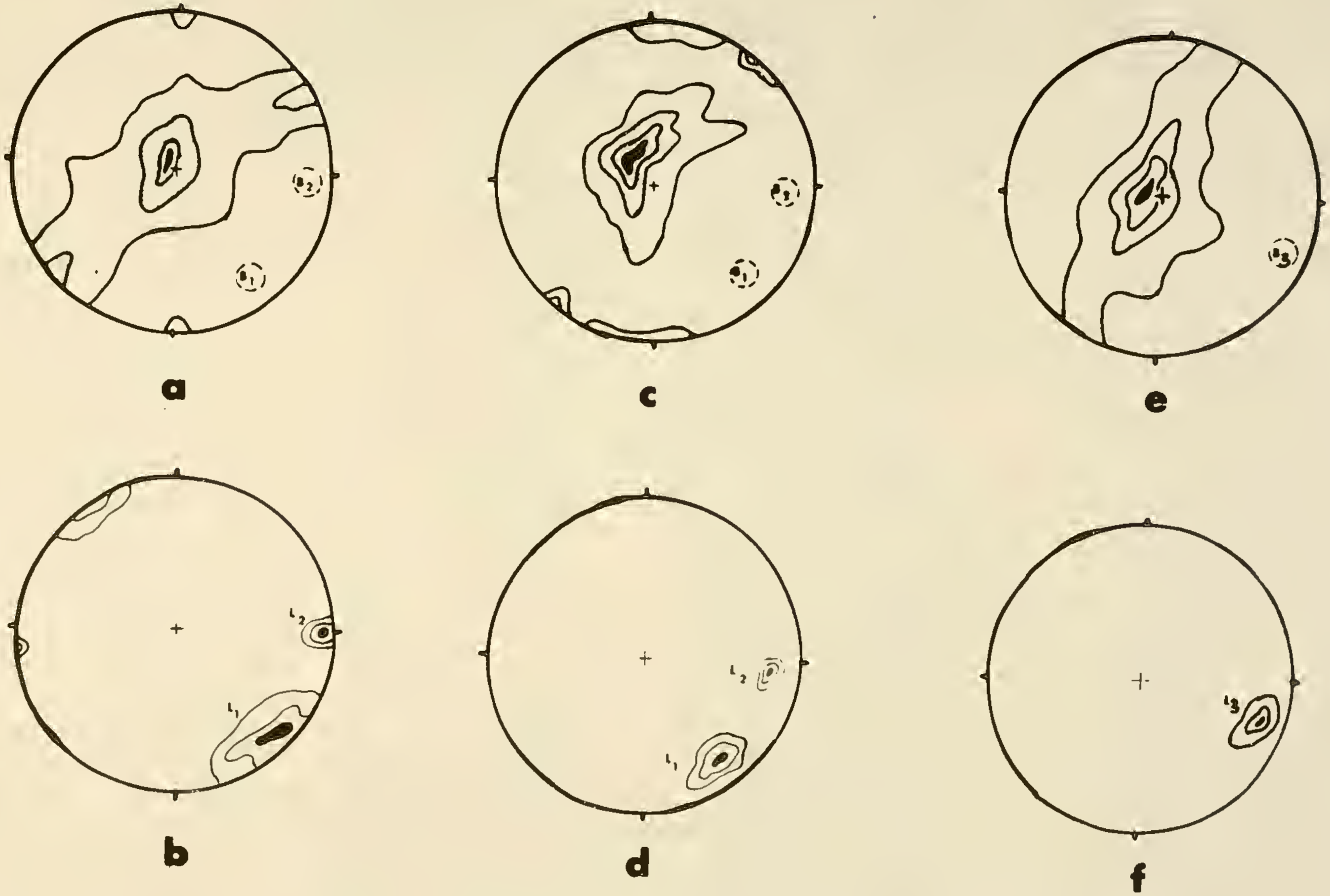

Fig. 4. Stereograms of foliations and lineations. a) and b) cover an area of $\sim$25 square miles between Peck Lake and Canada Lake. c) and d) cover an area of $\sim$25 square miles north and east of Peck Lake. e) and f) are for the Northville Syncline at the NE edge of the map.

B_1, B_2, B_3 and L_1, L_2, L_3 refer to fold axes and lineations of the associated F_1, F_2, and F_3 folds.

Plots of foliation (a), c, e)) are contoured at 1,3,5, and 7 percent of the area. Plots of lineation (c), d), f)) are contoured at 5, 10, and 15 percent of the area. Figures a), b), and c), contain between 200 and 300 points each. Figures c), d), and f) contain between 50 and 100 points each.

axis remains unsolved. This investigator favors the possibility that the axis swings north around the quartz monzonite block. However, it is difficult to substantiate this hypothesis. This is in large part due to the problems inherent in subdividing the biotite quartz-oligoclase gneiss. It is also due to the fact that the region at the northern extremity of the quartz monzonite block is grossly complicated by small bodies of diabase, metagabbro, and granitic rocks. The geology is further complicated by the Northville Syncline.

Ultimately, the location of the F_1 fold axis will be most convincingly shown by large scale mapping to the east. It is for this reason that the author is currently working in the adjoining Broadalbin 15' quadrangle. The resulting regional stratigraphy will very likely give a clear picture of the structure of the entire southern Adirondacks. Subsequently the problem of the extension of the F_1 fold axis around the quartz monzonite will take on secondary importance.

F_2 Fold System

The F_2 fold system is best represented by the large east plunging syncline whose axis trends N70W to E-W through the southwestern corner of the map. This fold plunges 10°-15° to the southeast. Its axial plane is warped, but, in general, dips north at steep angles. The topology of the F_2 fold is depicted in figure 4.

The style and attitude of the F_2 fold are well displayed by associated minor folds. A nice set of these are developed in the outcrop along Sprite Creek (stop I) and at the junction of Willie Road and Peck Hill Road (stop A). These folds tend to show sharp and fairly tight bends which open into more linear limbs, i.e.- they resemble Z's or angular S's. In this respect they exhibit shapes similar to the smaller folds that appear on the north limb of the F_1 fold in figure 2. In fact, these folds may be drag folds related to the larger F_2 fold. Their axial trends of N50W to E-W are consistent with this conclusion.

The most ubiquitous feature associated with the F_2 fold system is the N60W to E-W mineral lineation that is pervasive throughout the area. This lineation develops into rodding in many places. Such rodding is well developed in the roadcuts along Route 29A between Pine Lake and Stratford. In many of these exposures rodding has resulted in the development of pencil gneisses. Most of these pencil gneisses have had their foliation nearly obliterated. This is particularly true on surfaces perpendicular to the rod axes.

Possible F_3 Folds

At the southern and northern margin of the eastern quartz monzonite gneiss intrusive, there exist synclines that may represent locally developed F_3 folds. These folds are believed to have developed due to shouldering or fold interference by the quartz monzonite. It is possible that they are genetically related to the F_2 folds to which they are similar in style and orientation.

The best developed of the F_3 folds is the Northville Syncline whose axis trends N70W and plunges 15° east (fig. 3e) and f)). The axial plane dips about 45° to the south. Minor folds in the syncline exhibit sharp noses that are best characterized as resembling the letter, Z. Like the main fold, minor fold axial planes dip about 45° to the south. Strong mineral lineation and rodding are developed along the axis of this fold.

As shown on the map, the Northville Syncline does not continue to the west. It is largely for this reason that the structure is believed to be a local manifestation of structural influence due to the quartz monzonite block. The same is true of the more gentle syncline at the south margin of the quartz monzonite.

It is possible that the F_3 folds manifest responses of F_2 folding to local boundary conditions. Stereograms and the map pattern indicate that the quartz monzonite body has undergone some folding about an F_2 axis. The folds here described may represent F_2

strain taken up in the less rigid paragneisses.

F_4 and F_5 Fold Systems

There exist in the area broad open warps whose axes trend about E-W and N-S. Axial planes are vertical and axes are close to horizontal. These folds developed late in the history of the region and are of minor structural interest.

Possible and/or Slightly Reconnaissanced Folds

There is a possibility that an F_1 fold axis is associated with the very elongate E-W fold structure in the northern part of the area. The eastward closing of the quartzite and charnockite layers suggest this. Further to the east the leucogneisses thin down considerably and may close. This entire area has been examined for only a few days during July of 1969. Therefore, firm conclusions are not possible at this time.

a second possible fold structure is associated with the change of dips from 30° to vertical as one traces northern layers from east to west. This change of dip may be associated with the interaction of the rigid Piseco Dome block to the north and the thick quartz monzonite that crosses the west-central part of the area.

Faults

There exist a fairly large number of faults in the area. One of these defines the eastern margin of the area mapped in figure 2. Several other faults are shown in the southern part of the area. Aside from these instances, no other faults are shown on the map. This is because there are no known instances of large offset along faults within the area. For the present purposes they simply encumber small scale geologic maps.

The most widely developed set of faults trends N20E - N30E. This trend parallel the Piseco Fault. It is along these faults that breccias are developed. All of these breccias contain chlorite.

ROAD LOG

mileage

0 STOP A

Junction of Willie Road, Peck Hill Road, and Route 29A. Parking area just south of intersection.

Just to the north of Route 29A there are typical exposures of migmatite. Good F_1 minor folds may be found in most hand specimens. A few hundred feet up Willie Road there is a roadcut showing intense minor folding.

In this area strikes are NW and dips are to the NE. This is the structural pattern on the south limb of the large F_2 fold. F_2 minor folds are present.

This stop is close to the nose of the largest F_2 fold in the area (see figure 2).

Proceed northwest on Route 29A.

.3 Roadcuts on the south side of Route 29A exhibit the grey gneiss phase of the migmatite. Dips are vertical.

.6 100' back in the woods on the south side of Route 29A there are migmatite outcrops which exhibit excellent examples of F_2 minor folds.

1.2 Intensely folded migmatite on the north side of the road.

1.3 Mud Lake to the northeast of road.

2.8 Peck Lake to northeast side of road

3.6 STOP B

Roadcuts of garnetiferous quartz-biotite-oligoclase gneisses. These outcrops are typical of the unit. The overall attitude of foliation is N60W, 30°NE, but locally the dip goes to 70°NE or even vertical.

There is a fairly good development of minor folds in this outcrop. These trend around N50W and plunge 15° SE. Axial plane cleavage and lineation cut across compositional banding in some of these folds.

Some fault gouge is present in the outcrop and is associated with EW fractures. Also associated with these fractures are locally vertical dips.

The variety and complexity of the geology in this outcrop is typical of the quartz-biotite-oligoclase gneiss.

4.5 Roadcut of quartz-biotite-oligoclase gneiss on east side of road.

5.1 Turn East on Fisher Road.

mileage

5.6 STOP C
Southern contact between Royal Mt. Quartz Diorite and quartzitic units of the Ernst Corners leucogneiss. The contact is exposed about 50' south of the road. There are good examples of igneous features displayed in the quartz diorite. At the contact there are good quartzites.

Proceed back West on Fisher Road

6.1 Turn north on Route 29A.

7.1 Junction 29A - 10. Enter village of Caroga Lake

8.0 Junction 29A - 10 and Fulton County Road 112. Turn East.

9.1 STOP F
Small parking area on south side of road. Contact between Hogback Mountain Charnockites and Irving Pond Quartzites. The charnockites are exposed a few feet south of the road. The quartzites are exposed just north of the road.

Proceed back West

10.3 Junction 29 A - 10. Turn North.

11.2 Nick Stoner's Inn on west side of road.

11.6 STOP E
Parking area on west side of road. On the east side of the road there is an old logging road that goes east for about ¼ mile. At the end of this road proceed south up the hillside. On this hillside there are fine exposures of Irving Pond Quartzite. Interlayered with the quartzites are feldspathic bands and bands of amphibolite -pyroxenite. Some of the latter outline, or lie, in the cores of remarkable F_1 folds. It is in this area that the style and attitudes of F_1 minor folds are best developed. F_2 minor folds are also present.

11.9 Good exposures of pelitic members of the Irving Pond Quartzites. These occur right at the contact, which is exposed. Also present are unusually good F_2 minor folds with textbook examples of drag folding. A large boulder shows biotite lineation growing across compositional banding.

12.1 Large roadcuts of Hogback Mt. charnockites. From here to just north of Green Lake are the type exposures of this unit. Note the homogeneity and equigranular nature of the rock.

12.9 Cross contact into Green Lake Quartzite.

mileage

13.2 STOP D

Canada Lake Store and Post Office.

The large ledges across from the store are excellent examples of the Royal Mt. Quartz Diorite. As at stop C, these gneisses tend to be relatively homogenuous except for occassional amphibolite bands. Good igneous features are visible where the quartz diorite cross cuts, breaks up, and rotates an amphibolitic band at the east end of the outcrop.

Just beyond the visible top of the large ledge are good outcrops of garnetiferous quartz-biotite-oligoclase gneiss.

Walk east along the road between Green Lake to the north and Canada Lake to the south. Looking north note the rugged mountain known as Camelhump. The break between the two humps marks the contact between quartz diorite gneiss and quartz-biotite-oligoclase gneiss. Green Lake itself straddles the contact between quartz diorite gneiss and the Green Lake Quartzites. To the east, the steep hillside is composed of Hogback Mt. Charnockite with metagabbro at the top.

At the east shore of Green Lake enter the woods and observe a well exposed section of the Green Lake Quartzites. In this locality the quartzites are highly sheared.

Proceed up the hillside and cross the contact with the Hogback Mt. Charnockites. This is one of the best exposed contacts on the north limb of the F_1 fold.

14.4 Outcrops of quartz-biotite-oligoclase gneiss on north side of road.

15.1 Pine Lake. Junction 29A and Route 10. Proceed North along Route 10 towards Piseco Lake.

16.2 Northward dipping biotite-quartz-oligoclase gneiss on west side of road

17.9 East Stoner Lake to East. Crossing contact between biotite-quartz-oligoclase gneiss and the Rooster Hill Quartz Monzonite. Rooster Hill rises along the west side of the road.

18.5 New roadcuts of dark green weathering inequigranular quartz monzonite.

20.4 Outcrop of quartz-biotite-oligoclase gneiss inclusion (see figure 2).

mileage

20.8 STOP G
Good examples of fresh and weathered surfaces of inequigranular quartz monzonite. Megacrysts are clearly visible. On fresh surfaces they are green and on weathered surfaces they are white. Thin sections show these megacrysts to be orthoclase and microperthite. Thin section examination also shows the presence of 25-30% quartz in this rock. Mafics are hornblende and orthopyroxene.

Lineation of the K-feldspar megacrysts is particularly striking on the weathered surfaces.

21.2 Parking Area

21.4 Cross creek and re-enter quartz-biotite -oligoclase gneisses.

23.4 Strongly rodded biotite-quartz-oligoclase gneiss in roadcut.

24.1 Trail to Tomany Mt. ranger tower.

24.4 Leucocratic gneiss exposed along road.

24.6 Leucocratic gneiss and biotite-quartz-oligoclase gneiss in roadcuts. Intrusive relations are visible.

24.7 Avery's Hotel

25.1 Leucogneiss on east side of road.

25.3 Crossing contact between the leucogneiss and a basic phase of the Cathead Mt. charnockite.

25.8 Basic phase of charnockite

26.9 Cross Creek. Outcrop of good olivine metagabbro displaying relict ophitic structure.

27.5 Large roadcuts of Cathead Mt. Charnockite are on the west side of the road. To the east are swamplands. Note the "swaths" of pink charnockite that occur in the otherwise light green roadcut.

28.2 Shaker Place turnoff.
Quartzites and quartzitic paragneisses. This band bisects the Cathead Mt. Charnockite and can be traced eastward for a distance of 15 miles.

The quartzitic gneisses are steeply dipping and appear to be involved in the folding that is beautifully displayed by amphibolite bands.

mileage

29.0 Crossing contact with leucogneiss unit. Excellent, steeply dipping leucogneisses are exposed in the woods just east of the road.

Return to Pine Lake Junction with Route 29A.

44.0 Intersection of Routes 29A and 10. Turn west.

44.6 Old road metal quarry of dark quartz-biotite-oligoclase gneiss cut by pink pegmatite.

45.3 Road cut of quartz-biotite oligoclase gneiss.

46.3 Road cuts of steeply dipping and strongly rodded biotite-quartz-oligoclase gneiss. This outcrop contains inclusions of hard, green colored knots of diopside and sphene. Farther up the hill to the north is a leucogneiss band that dies out to the east and to the west.

48.3 East Canada Rod and Gun Club. On the north side of the road there is a small roadcut in the biotite-quartz-oligoclase gneiss. In this roadcut there is developed an excellent F_1 minor fold. The axis trends N70W and is nearly horizontal.

50.6 Large roadcuts of strongly rodded biotite-quartz-oligoclase gneiss. Between here and the last stop there are several good roadcuts of this gneiss. In this exposure foliation trends almost E-W and is close to vertical. The strong rodding tends to obliterate much of the foliation and on some surfaces it is totally obliterated.

52.2 Junction with N-S road. Turn south.

54.5 Junction with E-W road. Turn east.

55.1 Junction with Stewart's Landing Road.

Proceed down road to Stewart's Landing. At the end of the road is the Stewart's Landing dam across Sprite Creek.

STOP H

The banks along Sprite Creek show excellent exposures of almost pure white and rose quartzites. These are among the best exposures of quartzite anywhere in the area.

To the NW of the creek is the contact with the Hogback Mt. charnockites on the north limb of the F_1 fold. This is the same contact exposed in the large roadcuts at the eastend of Canada Lake (mile 11.9). The proximity of this contact is indicated by the presence of pelitic bands along the NW bank of the creek just north of the dam.

mileage

Around Stewart's Landing the rocks strike N70E and dip 20°S. Return to junction at mile 55.1. Turn left at junction.

56.1 Outcrops of Hogback Mt. Charnockites striking N-S and dipping east on the nose of the F_2 fold. The charnockites are leucocratic and mesoperthite roch here.

57.2 STOP I
Road crosses Sprite Creek.
This is the same creek that is dammed at Stewart's Landing. However quartzites are no longer exposed in the creek. Instead there are well developed and typical brown, pink, and white weathering Hogback Mt. Charnockites. These strike NS and dip to the east. Present in the stream are some good examples of F_2 minor folds.
Note the strong lineation (L_2) in these outcrops.

58.3 Crystal Lake to NE of road.

61.8 Exposures of Hogback Mt. Charnockite outcrop on the north side of the road. These are charnockites of the south limb of the F_1 fold.

62.5 Small roadcut of quartzites and feldspathic quartzites of the F_1 fold core. These dip steeply (45°-60°) and strike N45W.

63.9 Bridge across creek. Northwest striking charnockites are exposed in the creek bed. These are charnockites of the north limb of the F_1 fold.

64.5 Lassellsville. Junction Route 67. Turn east.

67.7 Quartzites and feldspathic quartzites of the Irving Pond unit are exposed along the north side of the road for the next half mile. Strikes are NW and dips average 25° to the NE. Good 10'-20' thick bands of pure quartzite are exposed along the road. Less than a quarter mile northeast the charnockite is encountered again.

69.8 Junction Routes 10-29.

END

ACKNOWLEDGEMENTS

This work has been supported by grants from the Colgate University Research Council, a Sloan Faculty Fellowship, and NSF grant GA-528. Special thanks go to Yngvar Isachsen for his help and encouragement.

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NOTE

After this manuscript was prepared, further field work showed that the large east plunging F_2 syncline can be traced past the town of Salisbury Center. This extends the fold axis another 7 miles to the west.

Miller's (1909) map of the Remsen 15' quadrangle suggests that the F_2 fold axis may be traced as far northwest as Forestport -a distance of 35 miles NW from the western margin of the map shown in figure 2.

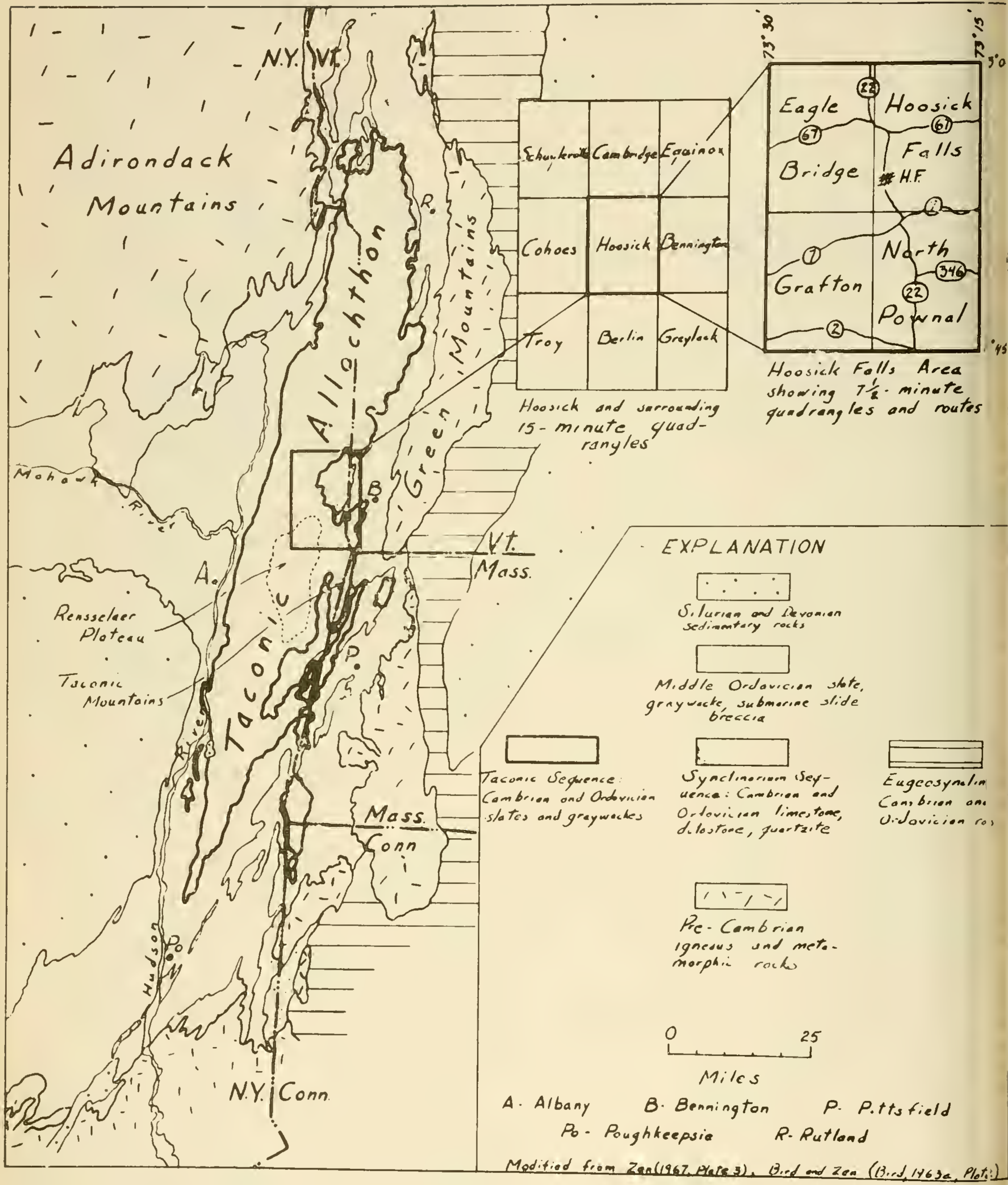

Figure 1. Geologic setting of the Taconic Allochthon and location of the Hoosick Falls Area

TRIP 12

SOME MAJOR STRUCTURAL FEATURES OF THE TACONIC ALLOCHTHON IN THE HOOSICK FALLS AREA, NEW YORK - VERMONT

by

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Hamilton College

Michael A. Lane
Indiana University

Purpose: We will see on this trip the two major thrust sheets that comprise the eastern part of the Taconic allochthon in this area. We will examine in detail some of the thrust contacts, and see the recumbently folded nature of the base of the lower thrust sheet. We will also see the Middle Ordovician submarine slide breccia, with its giant clasts, that occurs immediately beneath the allochthon.

Background and acknowledgements This trip is based on a ten-year detailed stratigraphic and structural study (Potter, in press) in the Hoosick Falls area (Figure 1). Field work has been supported by the New York State Geological Survey, the National Science Foundation, The Geological Society of America, and Hamilton College. Lane has made a detailed structural analysis at selected localities in the area for a PhD dissertation at the University of Indiana. His study is aimed at deciphering the deformation history and is not intended to be an assessment of thrust-no thrust problem.

E. Zen, W. Berry, J. Bird, G. Theokritoff, and D. Fisher have greatly aided Potter's study through field visits, identification of fossils, and through published data (see Zen, 1967 and references cited therein.)

Prior to the present work the most definitive study in the Hoosick Falls area was by Prindle and Knopf (1932). Bonham (1950), Balk (1953), and Lochman (1956) have also contributed to our knowledge of the geology and paleontology of this area. MacFadyen (1956), and Hewitt (1961) mapped the quadrangles east and northeast, respectively, of the Hoosick Falls area; and Metz (1969) has recently mapped the Cambridge Quadrangle to the north.

Stratigraphy While not the major concern of this trip, the stratigraphy of this area must be understood in at least summary fashion for the stratigraphic details enable us to establish structures which constitute prime evidence for the major thrusts. Figure 2 summarizes the relations of the two major stratigraphic sequences.

The Taconic Sequence, comprising the allochton, is approximately 4000 feet thick, and consists of turbidities and pelites suggesting deposition in deep water with unstable bottom conditions: delicately laminated argillite and thin-bedded chert suggest deep, quiet water conditions; euxinic conditions are suggested by pyritiferous black slate with and without graptolities; transportation and deposition by turbidity currents is indicated by the lithologic character, graded bedding and sole markings of the major graywacke units; unstable bottom conditions and submarine slumping are indicated by intraformational breccias (ibc, Figure 2) and by the presence of a few exotic clasts in some of the units. Stratigraphic units within the Taconic Sequence show great continuity north and south within the allochthon, but exhibit maximum change in thickness and in lithic character east-west (across strike).

Thus, practically every unit shown in Figure 2 can also be identified 60 miles north, at the north end of the allochthon.

The Synclinorium Sequence, largely synchronous with the Taconic Sequence, is about 2000 feet thick and consists of limestones, dolostones, and quartzites that attest to a shallow water shelf environment. These are overlain by slate-graywacke-submarine slide breccia that mark the period of Middle Ordovician thrusting.

Major structures and their evolution The allochthon in this area consists of two sheets, one above the other, that have been thrust westward onto the Synclinorium Sequence. Evidence for thrusting includes lithologic contrasts and gross structural discordance between synchronous formations above and below the thrust traces (Figure 3); slices of carbonate rock from the Synclinorium Sequence between the two thrust sheets (Stop 1); crushing, shearing, and mineralization at the thrust zones. The lower (North Petersburg) thrust sheet includes all the rocks of the Taconic Sequence except the Rensselaer Graywacke; recumbent folds are extensively developed in the lower 1000 feet of this sheet which consists of younger formations than the upper part of the sheet (structure sections, Figure 3). The North Petersburg sheet is thus a huge recumbent anticline or nappe (Figure 4), and it is correlated with Zen's (1967) Giddings Brook slice (Figure 5). Beneath the North Petersburg thrust is the Middle Ordovician Walloomsac formation consisting of slate, graywacke, and submarine slide breccia. The Whipstock submarine slide breccia contains clasts of the Taconic Sequence and some giant blocks of carbonate rocks from the Synclinorium Sequence. It is inferred that thrusting was a submarine phenomenon, that Austin Glen Graywacke was deposited on both Taconic and Synclinorium sequences at the early stages of orogeny, that as the thrust sheets moved into this area from the east, blocks of limestone and dolostone up to 1.8 miles long and 700 feet thick (from the shelf environment) and blocks of Taconic Sequence rocks (from the advancing thrust sheets) slid westward into the mud in the deeper parts of the basin to form the Whipstock. Unconsolidated breccia, graywacke, and mud were overridden by the North Petersburg sheet, and, because of the gross overturning of this sheet, unconsolidated Austin Glen Graywacke of the Taconic Sequence was locally melded with the unconsolidated material beneath the thrust.

The upper (Rensselaer Plateau) thrust sheet is perhaps the eastern core of the North Petersburg nappe which was thrust westward onto the core and inverted limb of the leading part of the nappe (Figure 4). On the plateau the Rensselaer Plateau sheet consists of Rensselaer graywacke and underlying Mettawee slate. Eight formations or stratigraphic units of the Taconic Sequence, ranging from Early Cambrian to Middle Ordovician, constitute the Rensselaer Plateau sheet on Mount Anthony and the Taconic Mountains. Identification of these units rules out MacFadyen's (1956) conclusion that the schists and related rocks here, which he called the "Mount Anthony Formation", are Middle to Upper (?) Ordovician and autochthonous. The correlation of this thrust sheet here with that capping the Rensselaer Plateau is based on the extensive exposures of Rensselaer Graywacke at the base of the sheet on Mount Anthony and on the Taconic Mountains (Figure 3), and on the fact that no other thrust sheet occurs between this one and the North Petersburg sheet or the autochthonous rocks below. Thus, Zen's (1967) Dorset Mountain slice in this area is considered to be the Rensselaer Plateau thrust sheet (Figure 5).

Both major thrust planes and thrust sheets have been refolded by a later stage of deformation that produced a pervasive slaty cleavage-foloation. All

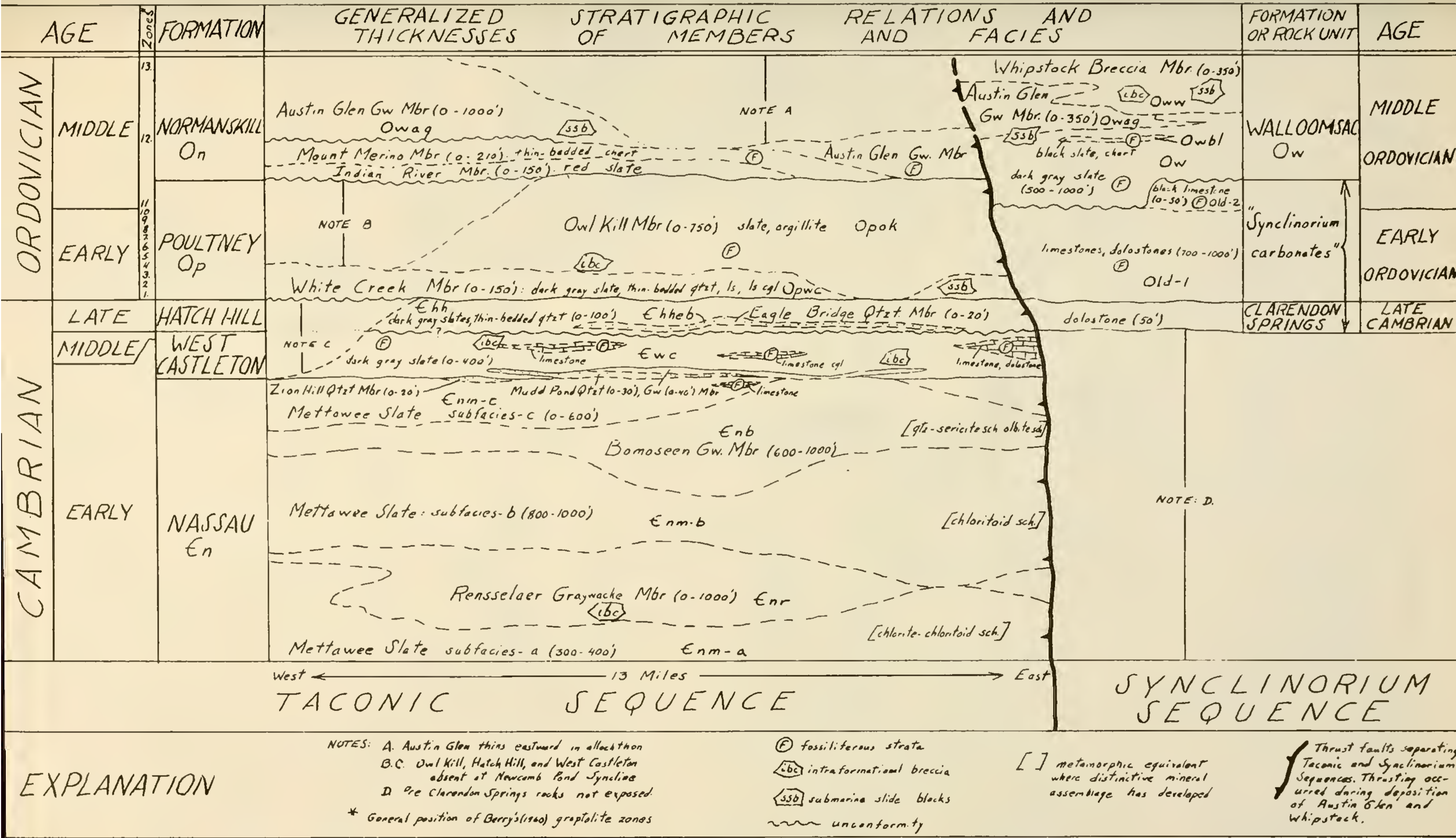
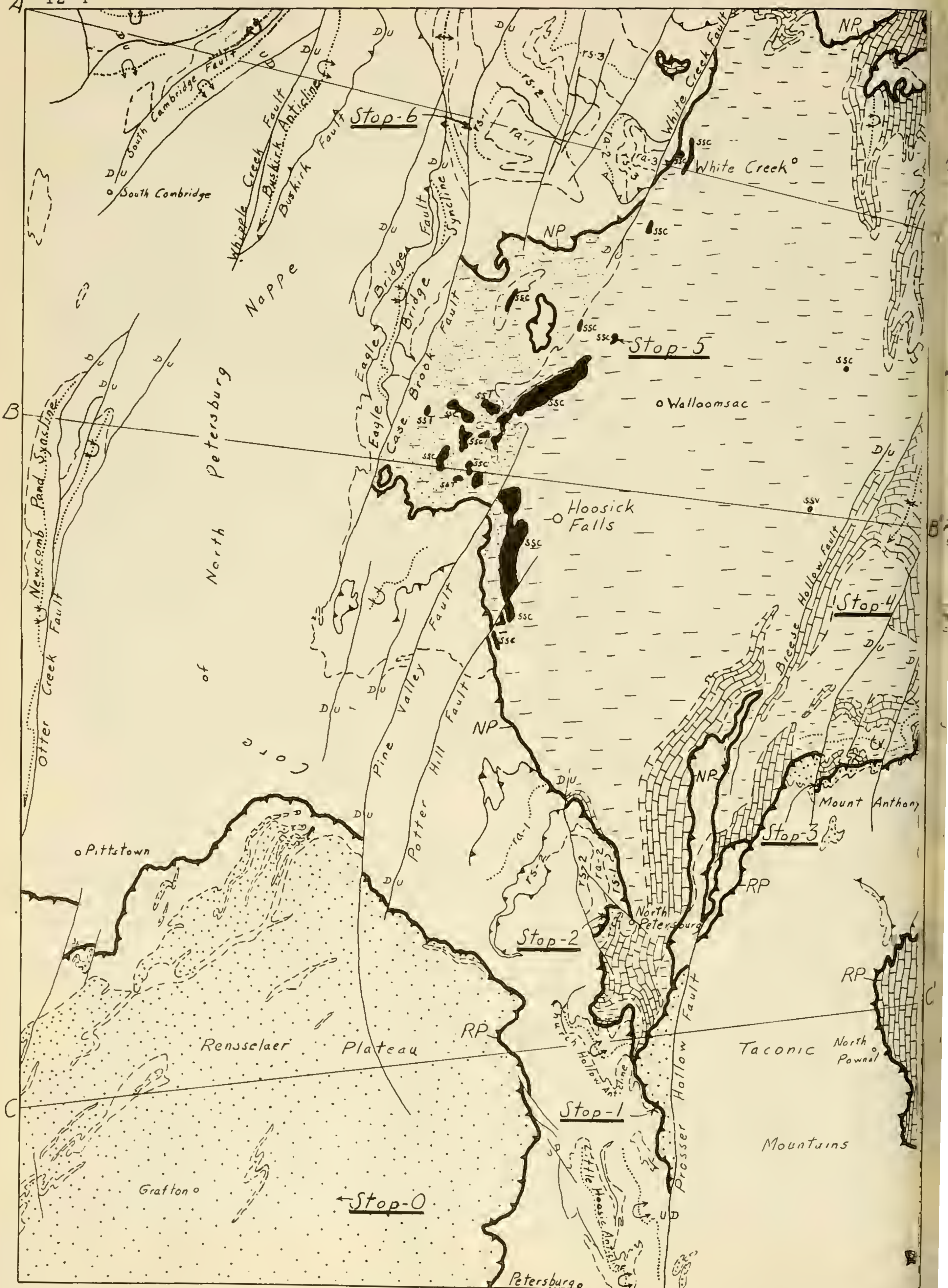


Figure 2. Stratigraphy of the Hoosick Falls Area


A
B
B'
C
C'
South Cambridge Fault
South Cambridge
Whipple Creek Fault
Buskirk Anticline
Buskirk Fault
Stop-6
White Creek Fault
White Creek
NP
Eagle Bridge Fault
Eagle Bridge Syncline
Case Brook Fault
North Petersburg Nappe
Stop-5
Walloomsac
Newcomb Pond Syncline
Otter Creek Fault
Hoosick Falls
Core of
Pine Valley Fault
Potter Hill Fault
Breese Hollow Fault
Stop-4
Mount Anthony
Stop-3
RP
Pittstown
North Petersburg
Stop-2
Rensselaer Plateau
Church Hollow Anticline
Prosser Hollow Fault
Taconic Mountains
North Pownal
Stop-1
Little Hoosic Anticline
Grafton
Stop-0
Petersburg

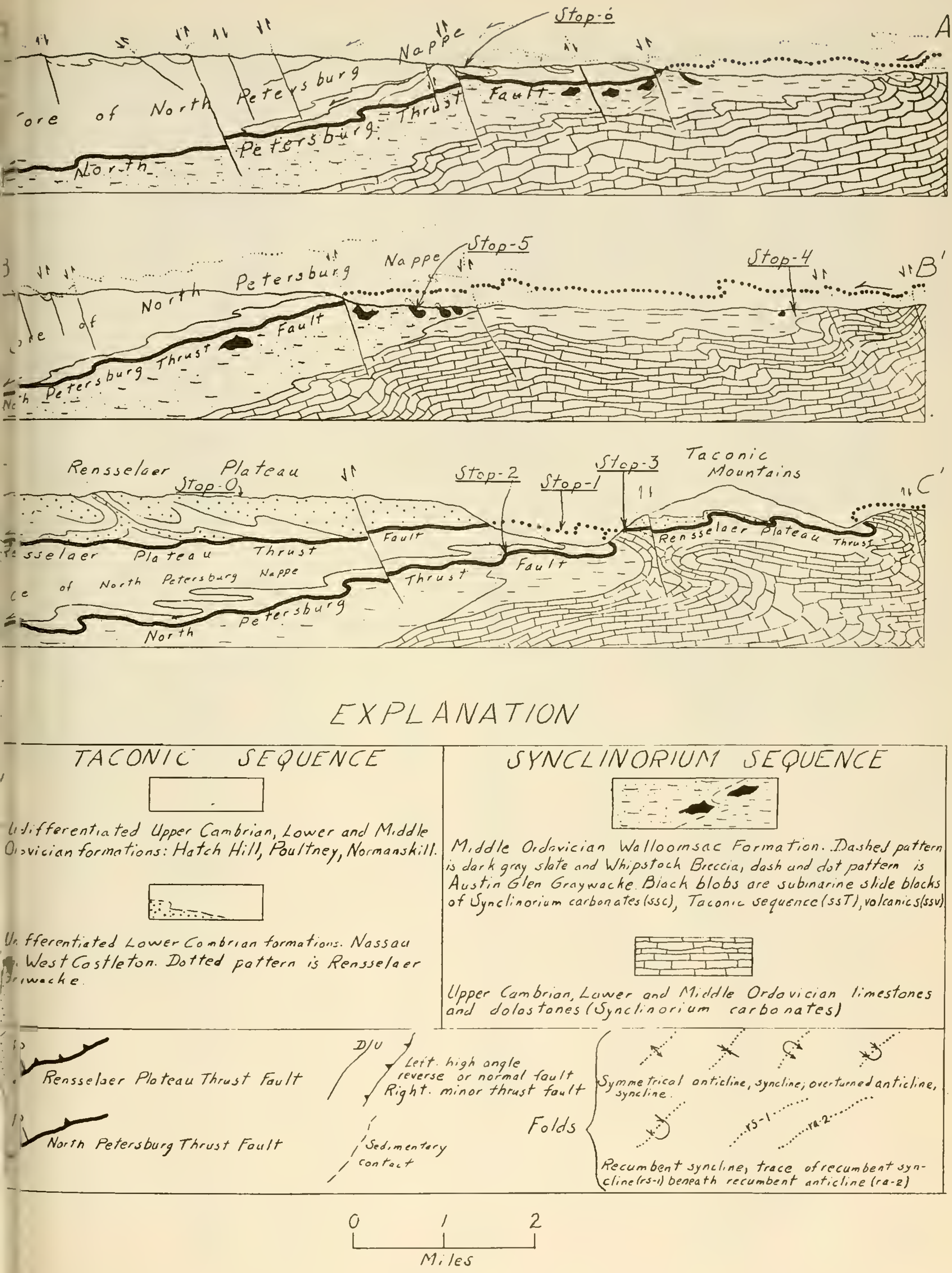


Figure 3. Geologic Map and Structure Sections of Hoosic Falls area

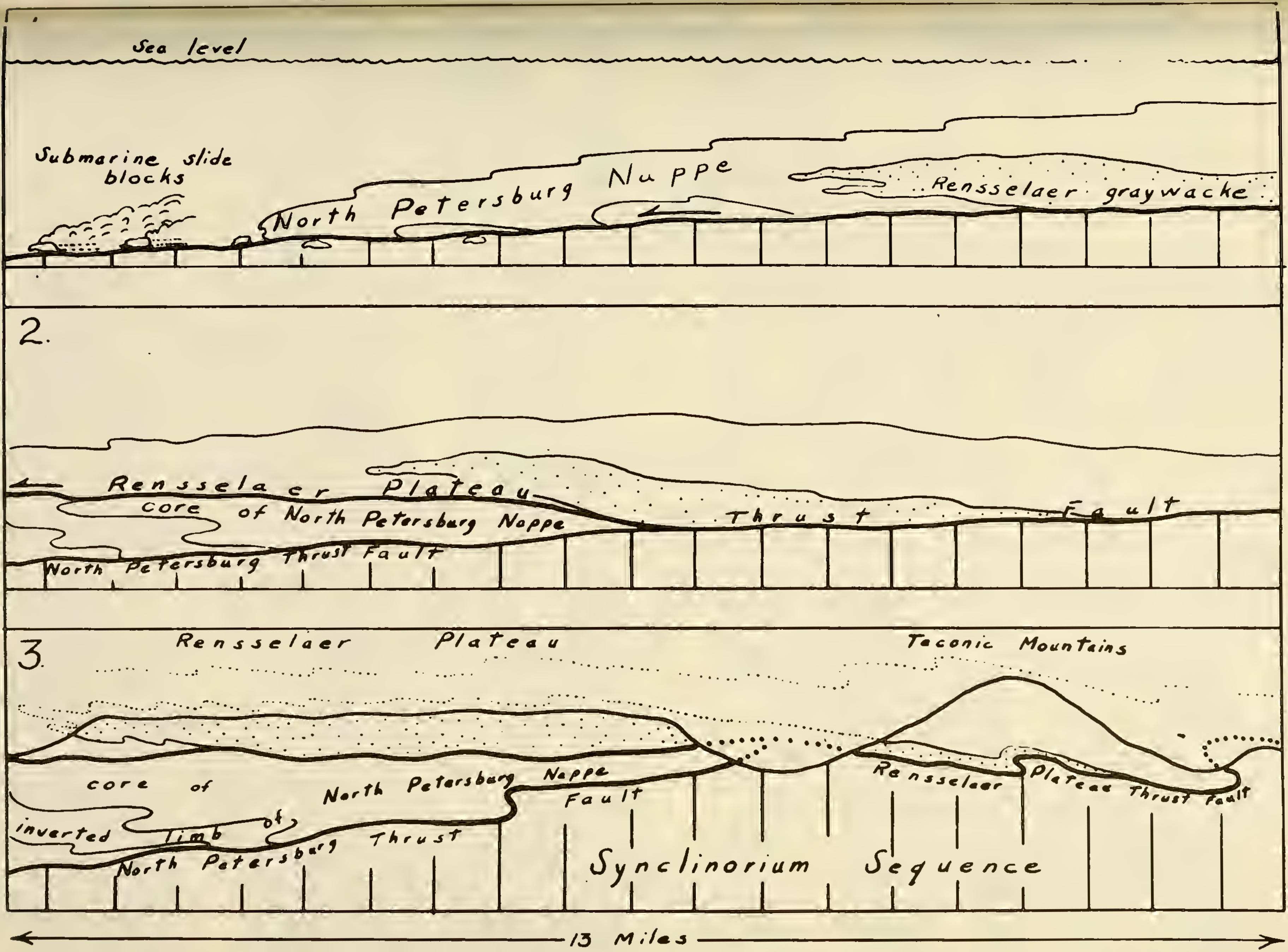
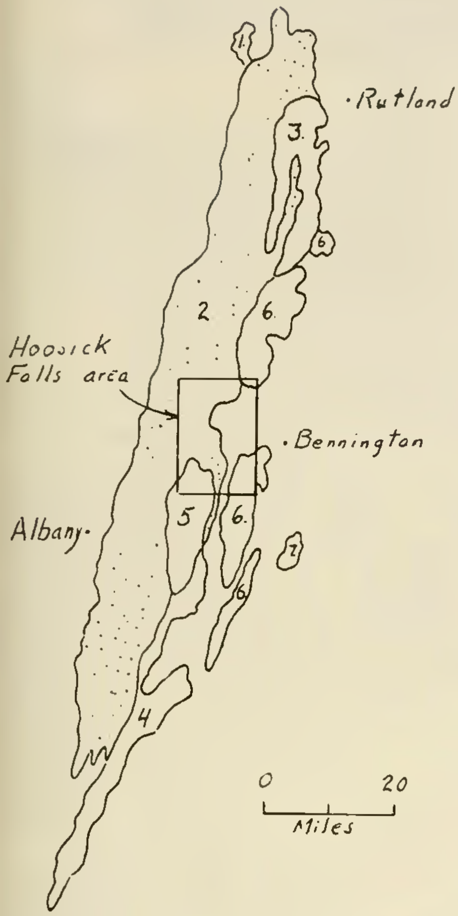


Figure 4. Schematic sections showing emplacement of thrust sheets



Thrust slices in Taconic Allochthon (abridged from Zen, 1967, Fig. 2)

Zen's (1967) thrust slices (in order of relative age of emplacement, oldest at bottom)	Thrust sheets in the Hoosick Falls Area (this report)
6 Dorset Mtn. and Greylock slices	Rensselaer Plateau sheet
5. Rensselaer Plateau slice	
4. Chatham slice	
3 Bird Mountain slice	
2. Giddings Brook slice	North Petersburg sheet
1. Sunset Lake slice	
	NOTE: Zen's slice 6 in south-eastern part of Hoosick Falls area is interpreted to be eastern half of Rensselaer Plateau sheet; in northeast corner of area, slice 6 is mapped as North Petersburg sheet.

Figure 5. Correlation of North Petersburg and Rensselaer Plateau thrust sheets with Zen's (1967) slices.

the rocks in the area underwent a regional metamorphism in Middle Ordovician time. Increase in rank from west to east is shown by the recrystallization of limestones and dolostones, and by metamorphism of argillites and slates to phyllites and schists containing chlorite, chloritoid, sericite, and albite. High-angle reverse and normal faults, striking north-northeast, cut the two thrust sheets and the autochthonous rocks beneath.

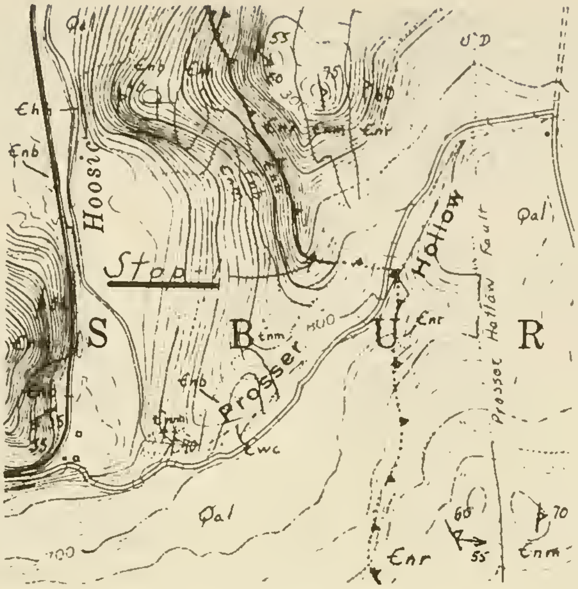
Lane's work suggests that four deformational episodes can be recognized in this area. The first, D_0, occurred at least in part before complete lithification of sediments, and consisted of large-scale westward transport and formation of recumbent folds and nappe structures. The next episode, D_1, produced a system of NNE-trending, westward-overturned folds and a pervasive axial plane slaty cleavage, S_1. Extensive mylonite zones along the Rensselaer Plateau thrust were formed contemperaneous with S_1, and metamorphism also occurred at this time. After the formation of S_1 and the mylonites, minor movement occurred on the Rensselaer Plateau thrust and perhaps on the North Petersburg thrust as well. D_0 and D_1 established the overall geometry of structures in the central Taconics. Later deformations, in this area at least, served only to modify the structure. During D_2, the pervasive slaty cleavage was refolded on a NNE axis, and an axial plane slip cleavage, S_2, was locally developed. The final episode of deformation, D_3, caused folding of S_1 about an ESE axis, and locally developed axial plane slip cleavage, S_3. The use of the term "deformational episode" is not intended to imply knowledge of temporally discrete deformations. It is possible that some of the deformations described may have been essentially continuous.

STOP DESCRIPTIONS

The location of each stop is plotted on the geologic map, Figure 3, and the general structural setting at each stop is indicated on the structure sections. The topographic maps (1:24,000) accompanying descriptions of Stops 1-6 show the limit of outcrops (fine dotted lines) and main geologic contacts. Refer to Figure 2 for letter symbols of stratigraphic units.

<u>STOP-0</u> Outcrops of Rensselaer Graywacke on north and south sides of Route 2, 0.2 mile west of East Grafton. These are good exposures of relatively unmetamorphosed medium and coarse grained Rensselaer Graywacke, an important stratigraphic unit on our trip. Note lack of slate interbeds between many of the graywacke beds. Graded bedding and intraformational clasts (phenoplasts) of dark gray siltstone are well shown in one bed on north side of road. The Rensselaer, in this area, is typically a feldspathic or lithic graywacke consisting of clasts of quartz, plagioclase, microcline, muscovite, cherty argillite, and graphitic quartzite in a dark green matrix of fine grained quartz, feldspar, chlorite, and sericite. (Collect a sample for comparison with the metamorphosed Rensselaer at Stop 3.)

STOP-1



North Pownal Quadrangle

Exposure of the Rensselaer Plateau thrust fault north of Prosser Hollow. Below the thrust fault is an apparently normal sequence of Bomoseen, Mettawee, and Hatch Hill (with Eagle Bridge Quartzite) - all part of the North Petersburg thrust sheet. The Rensselaer Plateau thrust fault is marked by a large sliver of limestone and dolostone of the Synclinorium carbonates that have been tectonically dragged to their present position. Immediately above the Rensselaer Plateau thrust is the Rensselaer Graywacke, perhaps several hundreds of feet thick and intensely sheared. The graywacke is faulted against chloritoid schist (Mettawee) 0.3 miles east of this stop.

The following details of the fault zone are noted. First, the Rensselaer Graywacke above the thrust is mylonitic through a zone approximately 150 feet thick (measured perpendicular to foliation), and the mylonitic foliation is concordant with normal foliation above and below the thrust zone (Figure 6). Second, the thrust plane truncates the mylonitic foliation. Third, a well-developed foliation parallel to the thrust plane occurs in the uppermost 2-3 feet of the limestone. Numerous other structural features may be observed. Widely spaced fractures parallel to the thrust plane also truncate the foliation and show a similar sense of movement to that on the thrust. Several warps in the thrust plane apparently represent areas where (later) movement on the thrust has locally followed the foliation instead of cutting across it. Near the upper (western) end of the outcrop, a sliver of mylonitic graywacke about 5' x 5' is completely enclosed within the limestone. West of this, the thrust plane steepens and follows the trend of the foliation in the graywacke for an indefinite distance.

Two generations of folds are occasionally visible in the mylonites above the mylonites above the thrust. In one generation, the axial planes are parallel to the foliation; the axes generally trend to the north but are variable. In some cases, the plunge of the axes is perpendicular to the strike of the axial plane, thus forming a reclined fold. This fold style is common in other thrust zones, notably along the Moine thrust in the Scottish Highlands. The second visible generation of folds has NNE trending axes and nearly vertical axial planes. These folds are correlated with F_2, one of the four fold systems in the non-mylonitic rocks in this area.

Interpretation: The earliest structural event well-represented at this stop is the formation of the pervasive axial plane foliation, S_1, and the accompanying regional metamorphism. Emplacement of the graywacke and chloritoid schist along the Rensselaer Plateau Thrust was apparently earlier than the formation of S_1. Prime evidence for this is the occurrence in several places along the thrust of tectonic slivers of autochthonous carbonates around which S_1 has been refracted.

The mylonites either were pre S_1 and rotated into their present orientation during the formation of S_1, or else formed at the same time as the foliation. Following the ideas of Johnson (1967) the latter explanation is preferred. Th mylonites are not necessarily related to large scale thrust movement, and consequently, evidence for emplacement of the Rensselaer Plateau thrust must come mainly from regional stratigraphic and structural studies.

Following S_1, minor movement occurred between the graywacke and the slates beneath. This movement caused the presently observed thrust plane, the thin zone of well-developed foliation in the upper few feet of the limestone, and the low angle fractures in the rocks immediately above and below the thrust. In other localities, notably west of the Little Hoosic Valley, and at Stop 3, the later movement caused a marked, but local, disturbance of foliation.

Explanation of Figure 6.

Equal area diagrams (lower hemisphere) showing orientation of poles to foliation in vicinity of Rensselaer Plateau thrust at Stops 1 (A-C) and 3 (D-F). Contours are 15%, 10% and 5% per 1% area unless otherwise indicated.

A. Below thrust at STOP-1: 105 measurements.

B. Thrust zone, within 5 feet of thrust plane, at STOP-1: 28 measurement

C. Above thrust at STOP-1: 42 measurements.

D. Thrust zone, within 5 feet of thrust plane at STOP-3: 44 measurements.

E. Above thrust, 20 to 50 feet vertically up slope to the NE of STOP-3: 20 measurements.

F. Regional trend of foliation in NE 4/9 of North Pownal Quadrangle. 410 measurements. Contours are 10%, 5% per 1% area.

These diagrams are intended only as a qualitative guide to foliation orientation. The contours are not statistically rigorous.

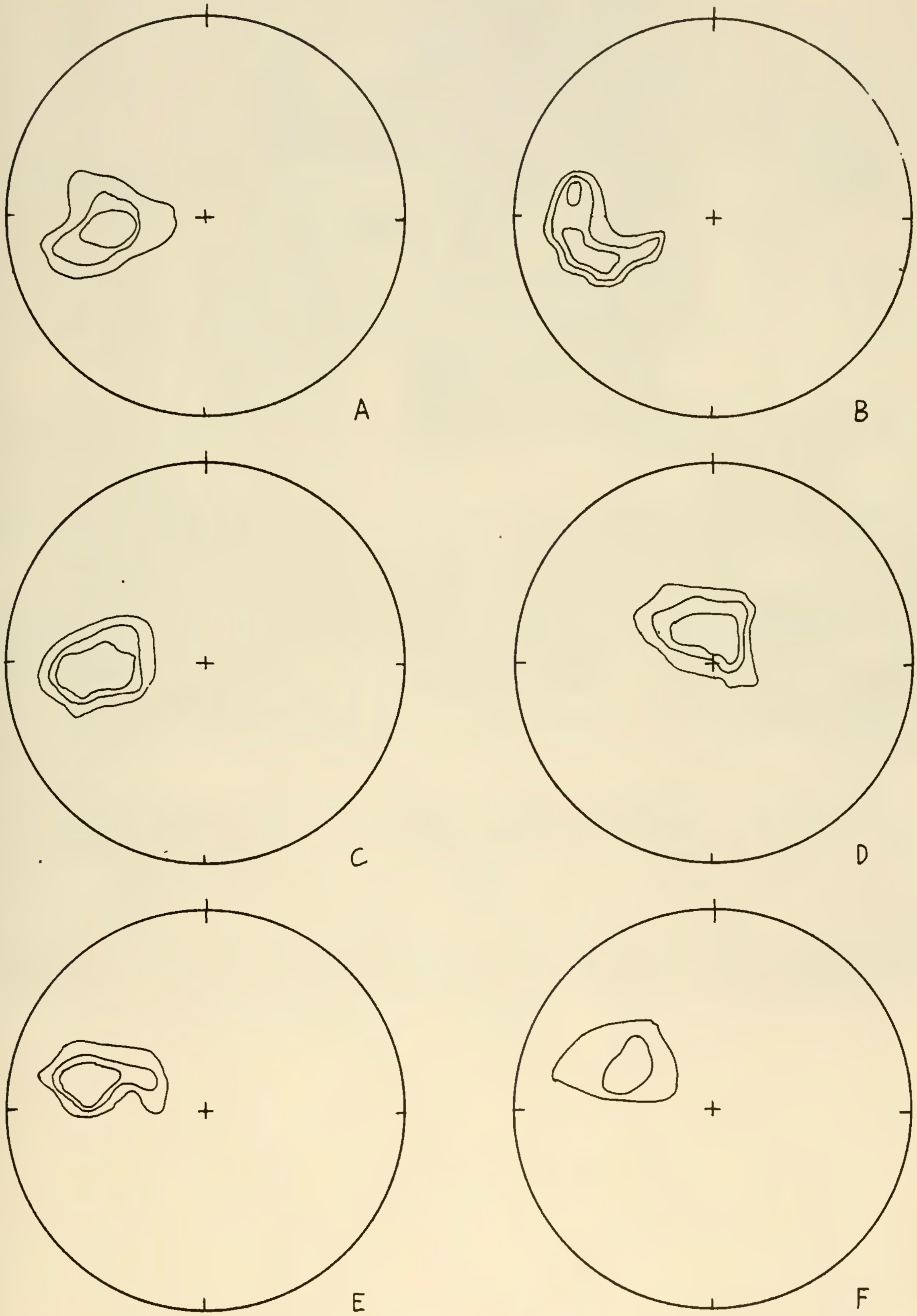
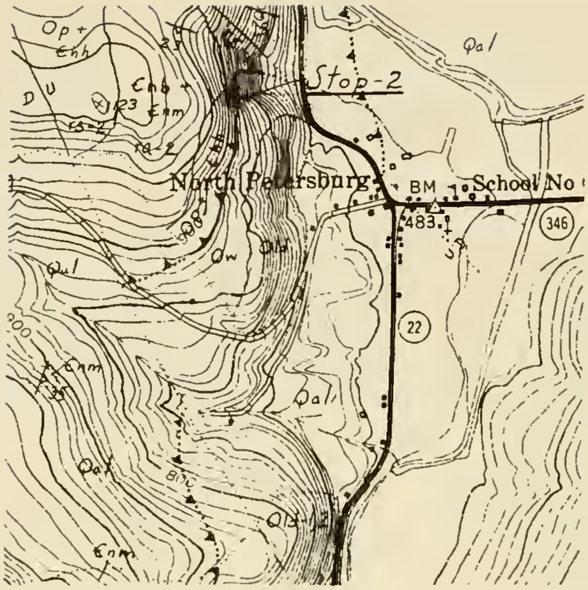


Figure 6 Equal area diagrams of poles to foliation
in vicinity of Rensselaer Plateau thrust fault.

STOP-2



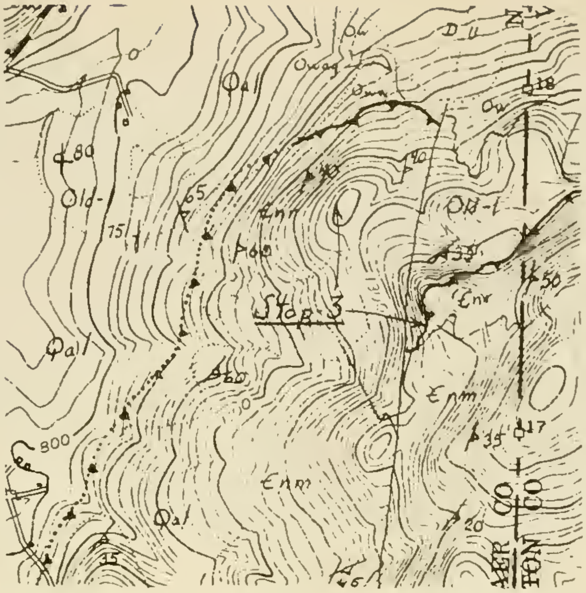
North Pownal Quadrangle

0 1000 Feet

Exposure of North Petersburg thrust fault, west of North Petersburg. Walk to thrust contact on steep slope (elevation 800 feet) via outcrops of limestone and dolostone of the Synclinorium carbonates. Steep slope above carbonates and below thrust is believed to be largely underlain by Walloomsac slate. There are a few small outcrops of slate on this slope, and a good exposure of the slate beneath the thrust fault 3/4 mile north of this stop. Less than five feet beneath the thrust fault, and inbedded in Walloomsac slate, is a large block of limestone (Synclinorium carbonate), interpreted to be a submarine slide block.

The thrust zone is characterized by shearing, mylonitization, bleaching and calcification of argillites, cherty argillites, and slates belonging to the Owl Kill Member of the Poultney Formation. Above this, through a vertical distance of some 200 feet, is an inverted sequence of the White Creek Member of the Poultney (ribbon limestones in black slate), Hatch Hill (thin-bedded quartzites in dark gray slate), Eagle Bridge Quartzite, and Bomoseen Graywacke. The Bomoseen marks the core of recumbent anticline, ra-2 (Figure 3), one of several recumbencies in this part of the area that characterize the lower part of the North Petersburg nappe.

STOP-3



North Pownal
Quadrangle

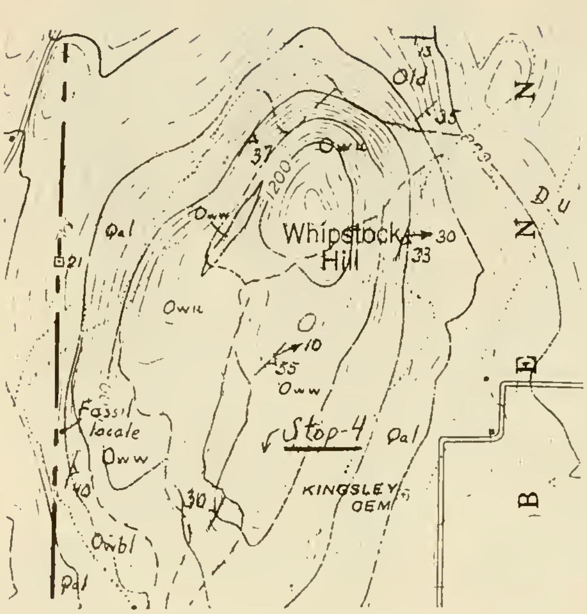
0 1000
Feet

Exposure of folded Rensselaer Plateau thrust fault on west shoulder of Mount Anthony (this stop treats you to a 3/4 mile ride or walk, 700 feet relief-one way). The rocks beneath the thrust are autochthonous Synclinorium carbonates-limestones and dolostones. In the low ground north of the shoulder of Mount Anthony these carbonates are recumbently folded (axis of fold trends east-west) with Walloomsac slate.

At closed contour 1300 we will see some least metamorphosed Rensselaer Graywacke. The thrust fault is exposed at elevation 1500. Immediately beneath the folded thrust plane the upper few feet of limestone is conspicuously thinly foliated, with foliation parallel to thrust plane. The graywacke above the thrust is schistese and contains characteristic seams of granular quartz. The foliation in the overlying graywacke is also parallel to the thrust plane, but within approximately 20-30' vertically, S_1 strikes NNE and dips moderately to steeply ESE, ie., concordant with the regional attitude of foliation (Figure 6). The graywacke is mylonitic for several hundred feet up the slope. In thin section and occasionally in outcrop, a slip cleavage is seen to displace the mylonitic foliation. This cleavage is not related to any of the regional fold systems.

As in Stop 2, the graywacke is believed to have been emplaced prior to the formation of the pervasive slaty cleavage and regional metamorphism. The mylonite formed at the same time as S_1, and was locally deformed by later movement on the Rensselaer Plateau thrust.

STOP-4



Hoosick Falls Quadrangle

0 1000
Feet

Exposure of Whipstock Breccia at Whipstock Hill. Several small scattered outcrops of Whipstock Breccia can be seen in the grassy terrain at the 1180 crest of Whipstock Hill and these afforded a close examination of the dark gray silty argillite or slate matrix, intraformational clasts, and a few exotic submarine slide blocks.

The matrix of the breccia is irregularily cleaved because of the intraformational clasts which range from about 1 mm. to 5 cm. in maximum dimension. Many of these small clasts lie in the plane of foliation and are smeared out and elongated so that they define a prominent lineation. The dominant clasts are fine grained serictic siltstone, green-gray argillite, quartzite, and fine grained limestone. The similarities between these lithologies and thin laminae and layers in the Walloomsac slate suggests an intrabasinal origin for the clasts. In addition, however, there are in the Whipstock exotic blocks of the Taconic Sequence, the Synclinorium carbonates, and volcanic rocks. Two exotic blocks can be seen at Whipstock hill: one is a conglomeratic quartzite (Rensselaer Graywacke or Zion Hill Quartzite), about 16 inches in maximum dimension, consisting of subrounded grains of quartz, quartzite, and oligeclase in a matrix of fine grained quartz, sericite, and chlorite. A second exotic block, about 4 x 4 feet in plan, is inequigranular pyritic quartzite resembling the Mudd Pond Quartzite.

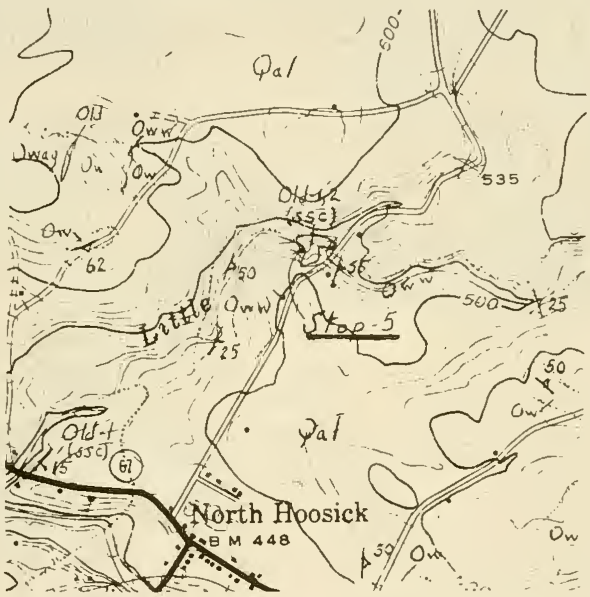
The Whipstock Breccia on Whipstock Hill is infolded with a large mass of fine grained phyllitic siltstone, and recrystallized radiolarian chert (Owu), rocks of unknown stratigraphic position. These rocks may be beds in the breccia, large intraformational clasts, or perhaps giant clasts of the Taconic Sequence.

The Whipstock is an integral part of the Walloomsac Formation. It is widely distributed beneath the North Petersburg thrust fault, and is closely associated with lenses of Austin Glen Graywacke. Its age is Wilderness or post-Wilderness for the breccia is underlain on the west slope of Whipstock Hill by a black slate containing graptolites of the Climacograptus bicornis Zone.

A similar submarine slide breccia (Forbes Hill) has been identified by Zen (1967) in the northern Taconics. Bird is now working on these "Wildflysch-type" conglometartes which are extensively exposed at the west edge of the allochthon (see Bird, 1963, pp. 17-19).

The crest and west flank of the Green Mountains are visible northeast of Whipstock Hill. Pre-Cambrian gneisses of the Mt. Holly Complex are exposed along the crest of the range, and the Lower Cambrian Cheshire Quartzite, which rests unconformably on the Mt. Holly, forms prominent dip slopes on the mountain flank. The Cheshire here is at the base of the Synclinorium Sequence. The Taconic Allochthon, being intermediate in facies between that of East Vermont (eugeosynclinal) and the Synclinorium Sequence (miogeesynclinal) presumably came from an area east of the exposures of the Cheshire Quartzite. Zen (1967) has proposed the Green Mountain core as the likely root zone. The prominent hill to the southeast is Mount Anthony, and the base of its steep north face marks the trace of the Rensselaer Plateau thrust fault.

STOP-5



Hoosick Falls Quadrangle

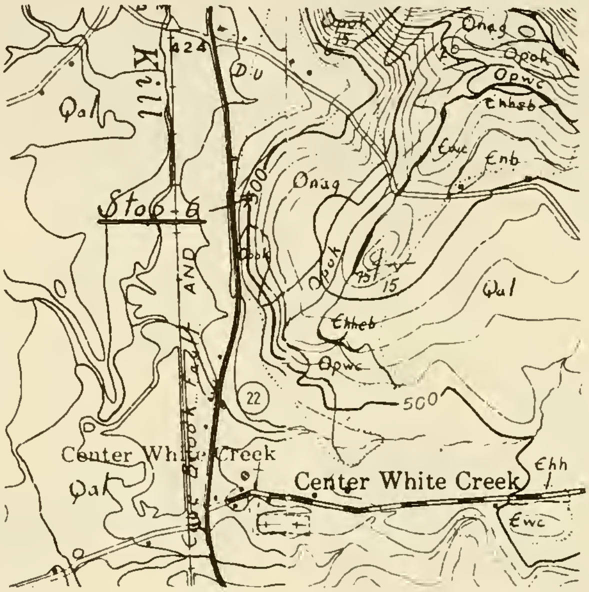
0 1000

Feet

Exposure of large submarine slide block in bed of Little White Creek. The block here consists of Synclinorium carbonates and measures about 300 x 400 feet in plan. Whipstock Breccia is exposed above and below the block which consists of highly folded limestone and dolostone (Old-1) overlain by black argillaceous limestone (Old-2). The contact between the black limestone of the block and the overlying Whipstock is locally conformable, but at the west (downstream) end of the exposure the contact between layers in the block and foliation in the breccia is discordant, and the Whipstock Breccia at the contact is crumpled, sheared, and faulted. Slaty cleavage in the Whipstock near the block is cut by a younger slip cleavage.

The problem of identity of masses of Synclinorium carbonates as discrete submarine slide blocks ("ssb" on Figure 3) is great for exposures in the western part of the Hoosick Falls re-entrant are poor, and these carbonate masses could be isolated outcrops of the complexly folded carbonate sequence. The following evidence suggests a submarine slide block origin: 1. an apparent structural discordance occurs from one block to another in the area southwest of STOP 5; 2. discrete masses (blocks) of the Taconic Sequence occur in the same terrain; 3. limestones at five of the "blocks" carry fossils of Early and Middle Ordovician age, yet the blocks are surrounded by Mid. Ordovician breccia, graywacke, or slate, or occur at the contact between graywacke and slate. The last relation suggests that submarine sliding occurred after a thick accumulation of mud (Walloomsac slate) and at the onset of deposition of the Austin Glen Graywacke.

STOP-6



Eagle Bridge and Hoosick Falls Quadrangles

0 1000

Feet

Recumbently folded Austin Glen Graywacke at roadcut on Route 22. The Austin Glen Graywacke, with interbedded dark gray slate and cross-bedded siltstones, is recumbently folded here in the core of recumbent syncline, rs-1, near the base of the North Petersburg thrust sheet. The exposure is apparently just east of the Case Brook reverse fault. Axial planes of recumbent folds here strike between N25° E and N80° E and dip from 13° to 23° southeast; axes plunge east-southeast from 3 to 20 degrees. Recumbent anticlines open to the northeast, synclines open to the southwest.

The exposure is notable for its wealth of primary sedimentary structures which includes subtle graded bedding and sole markings in the graywacke beds, and various types of cross-bedding in the siltstones.

Trip 12: Mileage Log

Leave Thruway Motor Inn near Exit 24 and proceed to intersection of Routes 2 and 7 in Troy. Mileage starts at this intersection.

<u>Mileage</u>

00.0	Intersection of Routes 2 and 7 in Troy. As we climbed the steep hill on the east side of Troy we crossed the west edge of the Taconic allochthon (Logan's Line), marked by Normanskill Formation with chaotic structure. Our trip , eastward on Route 2 from this intersection,takes us across the allochthon to its east edge.
04.3	Eagle Mills
06.6	Intersection of Route 278 with Route 2. Our route has been entirely on the lowest thrust slice of the allochthon (Zen's Giddings Brook slice, Figure 5). To the east we can see the west edge of the Rensselaer Plateau, and at its base is the trace of the Rensselaer Plateau thrust fault.
07.0	Road to Poestenkill and Brunswick goes off to right (south).
07.6	Cropseyville Post Office.
08.1	Cross approximate trace of Rensselaer Plateau thrust fault and continue eastward on the Rensselaer Plateau thrust sheet.
08.5	General store on left, exposure of Rensselaer Graywacke on north side of road two hundred feet north of store.
08.8	Rensselaer Graywacke on north side of highway.
09.6	Rensselaer Graywacke on south side of highway.
10.3	At bend in Route 2, road to left leads to Quackenkill quarry where Rensselaer Graywacke is quarried for road metal. There are excellent exposures of massive graywacke in abandoned quarries reached from this side road. Just east of quarry road, on north side of Rt. 2, are crops of Mettawee slate and Rensselaer Graywacke.
11.5	West edge of Hoosick Falls area (west edge of Grafton Quadrangle).
14.0 - 14.2	Grafton
15.4	School on north side of Route 2; road to Babcock Lake goes off to north.
15.9	Outcrops of Rensselaer Graywacke both sides of highway.
16.1	STOP-0: BEWARE OF TRAFFIC. Road cuts in Rensselaer Graywacke on both sides of road.
16.3	East Grafton

Mileage	
17.5	Rensselaer Graywacke on south side of highway.
18.5	Rensselaer Graywacke on south side of highway.
18.9	Rensselaer Graywacke on south side of highway. Driveway on north side of highway has cobblestone posts. Trace of Rensselaer Plateau thrust fault about 100 feet east of graywacke outcrops. Our route now continues eastward on the lower (North Petersburg) thrust sheet.
20.2	Intersection (underpass) of Route 22 with Route 2, Petersburg. Turn north on Route 22.
20.4 - 20.6	Mettawee slate: subfacies -b forms massive exposures on west side of highway. Moon Hill (at eastern digitation of Little Hoosic Anticline, Figure 3) is the prominent hill on the east side of the valley.
20.8 - 21.0	Bomoseen Graywacke and Mettawee slate (subfacies -b) exposed on west side of highway.
21.5	Cross Dill Creek
21.8	Odell Hill across valley to east. This hill is at north end of eastern digitation of Little Hoosic Anticline, Figure 3.
22.5	South end of massive exposures of Mettawee slate (subfacies -b), marking the core of the North Petersburg nappe. Turn right off Rt. 22 on Prosser Hollow Road.
22.7	Cross Little Hoosic River
23.4	White house on south side of highway, barn on north side. Debark for STOP-1. Walk up (west) across field to spur for exposure of Rensselaer Plateau thrust fault.
24.3	Junction of Prosser Hollow Road and Route 22; turn north on Route 22.
24.3 - 24.8	Massive exposures of Mettawee slate (subfacies -b) on west side of highway.
25.0	Bomoseen Graywacke on west side of highway.
25.6	Barn on east side of highway, house on west. We are at north edge of younger (shaded) formations at core of Church Hollow anticline (Figure 3). Bold cliffs on Taconic Mountains to east are Rensselaer Graywacke near base of Rensselaer Plateau thrust sheet.
25.7	Cross trace of North Petersburg thrust fault, and proceed to North Petersburg over autochthonous Synclinorium carbonates and Walloomsac slate.
26.5 - 26.6	Large exposures of recumbently folded Synclinorium carbonates on west side of highway.
27.3	Intersection of Route 346 with Route 22, North Petersburg. Turn west up hill. Steep slope marks exposures of Synclinorium carbonates. Cross trace of North Petersburg thrust fault at top of steep slope.

Mileage

27.9 Debark at farmhouse on north side of road for STOP-2. 20 minute walk involving steep slate-mantled slopes to see Synclinorium carbonates and the North Petersburg thrust zone.

28.5 Intersection of Routes 22 and 346 at North Petersburg. Proceed east on 346.

28.8 Cross Little Hoosic River.

29.1 Turn north off 346 at intersection. Grassy low hills ahead underlain by Synclinorium carbonates; approximate trace of North Petersburg thrust fault marked by lower edge of woods.

29.3 Cross B&M Railway and Hoosic River, bear left at intersection.

29.5 Cross 10-ton-limit bridge over B&M Railway. Immediately north of bridge are outcrops of Synclinorium carbonates.

29.7 Walloomsac slate on right.

30.4 Turn right at road intersection and proceed around south end of Indian Hill (long finger of North Petersburg thrust sheet, Figure 3) on County Road 20. View to south (right) into Little Hoosic Valley with Rensselaer Plateau on west side of valley and Taconic Mountains on east.

31.0 Slates of the Owl Kill Member of the Poultney Formation on west side of road.

31.8 - 31.9 Cross sheared and contorted sliver of Synclinorium carbonates which marks the hanging wall of Breese Hollow reverse fault. Walloomsac slate on foot wall to west.

32.4 Turn right (east) off County Road 20, and proceed to Cipperly farm.

32.7 Debark at farmyard for STOP-3: Good exposure of metamorphosed Rensselaer Graywacke in R. P. thrust sheet, and of folded R. P. thrust plane.

33.0 Intersection of County Road 20 and Cipperly farm road. Proceed north on County Road 20. North nose of Mount Anthony visible to northeast. Rensselaer Plateau thrust fault is at base of upper steep slope. Green Mountains visible in background to north.

33.3 Walloomsac slate on west side of road.

33.7 Walloomsac slate on west side of road.

34.3 - 34.4 Walloomsac slate and Whipstock Breccia on west side of road.

34.8 Red barn on right (north) side of road, Synclinorium carbonates in field on south side.

34.9 Intersection of County Road 20 and NY-7. Turn east on Route 7.

Mileage	
35.5	New York-Vermont line. Start Vt. -9.
36.4	Turn left (north) off Vt. -9 on Houran Road.
36.9	Debark at road bend for STOP-4: Whipstock Breccia on crest of Whipstock Hill.
37.4	Intersection of Houran Road and Vt. -9. Turn right on 9.
38.3	New York-Vermont line.
39.7	Turn right (north) off N.Y. -7 on East Hoosick Road (County Road 51). Our route now takes us into the western part of the Hoosick Falls re-entrant.
40.3	Walloomsac slate on left side of road.
40.8	Y intersection, bear left.
41.5	Road from north intersects County Road 51. Keep straight.
42.8	Walloomsac slate on right side of road.
43.0	Walloomsac slate on right side of road.
43.1	Keep straight at intersection on County Road 124. County Road 51 bears left.
43.4	View of west edge of Hoosic Falls re-entrant. The base of the hills to the west-across the valley of the Hoosick River-marks the approximate trace of the North Petersburg thrust fault.
44.6	Intersection of County Road 124 and Rt. 22. Turn right (north) on 22.
45.1	Cross bridge over Walloomsac River.
45.3	Intersection of Rts. 22 and 67. Turn left on 22.
45.35	Turn right off Rt. 22 on White Creek Road.
45.9	Debark for STOP-5 at private parking area near Little White Creek. USE NO PICKS AT THIS STOP. BE CAREFUL NOT TO DAMAGE DECKS OR WALKWAYS. FOLLOW THE LEADER. We will see a large submarine slide block of Synclinorium carbonates surrounded by Whipstock Breccia and Walloomsac slate.
45.95	Continue north on White Creek road. Cross bridge over Little White Creek. Whipstock Breccia in stream bed upstream from bridge (right side of road). Some clasts (not the cobbles in old concrete dam) in breccia are 6 to 8 inches in diameter.
46.4	Y intersection. Bear left on dirt road, then straight ahead at intersection 150 feet north.

Mileage

47.7 Turn left (west) at intersection on County Road 63. Recumbently folded formations at base of North Petersburg nappe exposed on steep wooded hill to north.

48.2 County Road 63 intersects with road leading south to Eagle Bridge. Keep straight on 63. Grassy hill in foreground to south is underlain by fossiliferous West Castleton, at the base of the North Petersburg nappe. Low grassy land south of hill underlain by Austin Glen Graywacke Member of the Walloomsac. In middle distance to south is the North Hoosick klippe with trace of North Petersburg thrust fault at lower edge of woods, Austin Glen beneath the thrust, and allochthonous Lower Cambrian formations above.

48.5 Intersection of County Road 63 and Delevan Road. Grassy hill on right (north) capped by fossiliferous West Castleton limestone, and dolostone.

49.3 Intersection of Lincoln Hill road with County Road 63 at Post Corners.

49.6 Recumbent anticline (ra-1, Figure 3) in low ground to right (north), nested below other recumbent folds which are well exposed on slopes of hills in background.

49.8 Hatch Hill black slate with interbedded calcareous quartzites on right side of road.

50.5 Intersection of County Road 63 and Rt. 22. Turn right (north) on 22.

50.8 Slate of the Owl Kill Member of Poultney Formation exposed on east side of highway.

50.9 Debark for STOP-6. BEWARD OF TRAFFIC. We will see here recumbently folded Austin Glen Graywacke Member of the Normanskill, near the base of the North Petersburg thrust sheet. Gross structures best seen from west side of highway.

50.9 Depart STOP-6. Go south on 22.

52.9 Intersection of Routes 22 and 67. Bear right (west) on 67.

53.3 Bridge over Owl Kill.

53.8 Bridge over Hoosic River.

54.0 Village of Eagle Bridge.

56.1 General store at Buskirk.

56.2 Small outcrop of Mettawee slate (subfacies -b) on north side of road, just east of underpass.

59.5 Cross Nipmoose Brook.

Mileage

59.8 Pass under B&M Railway

60.0 Two dirt roads intersect 67 from east and south. Normanskill Formatior on west limb of Newcomb Pond Syncline, approximately at Section B-B', Figure 3, is exposed 400 feet east of this intersection.

60.4 West edge of Hoosick Falls area (west edge of Eagle Bridge Quadrangle

END ROAD LOG

Return to Albany via Route 67 to Schaghticoke, and Route 40 through Grant Hollow (Deep Kill) and Diamond Hill to Troy.

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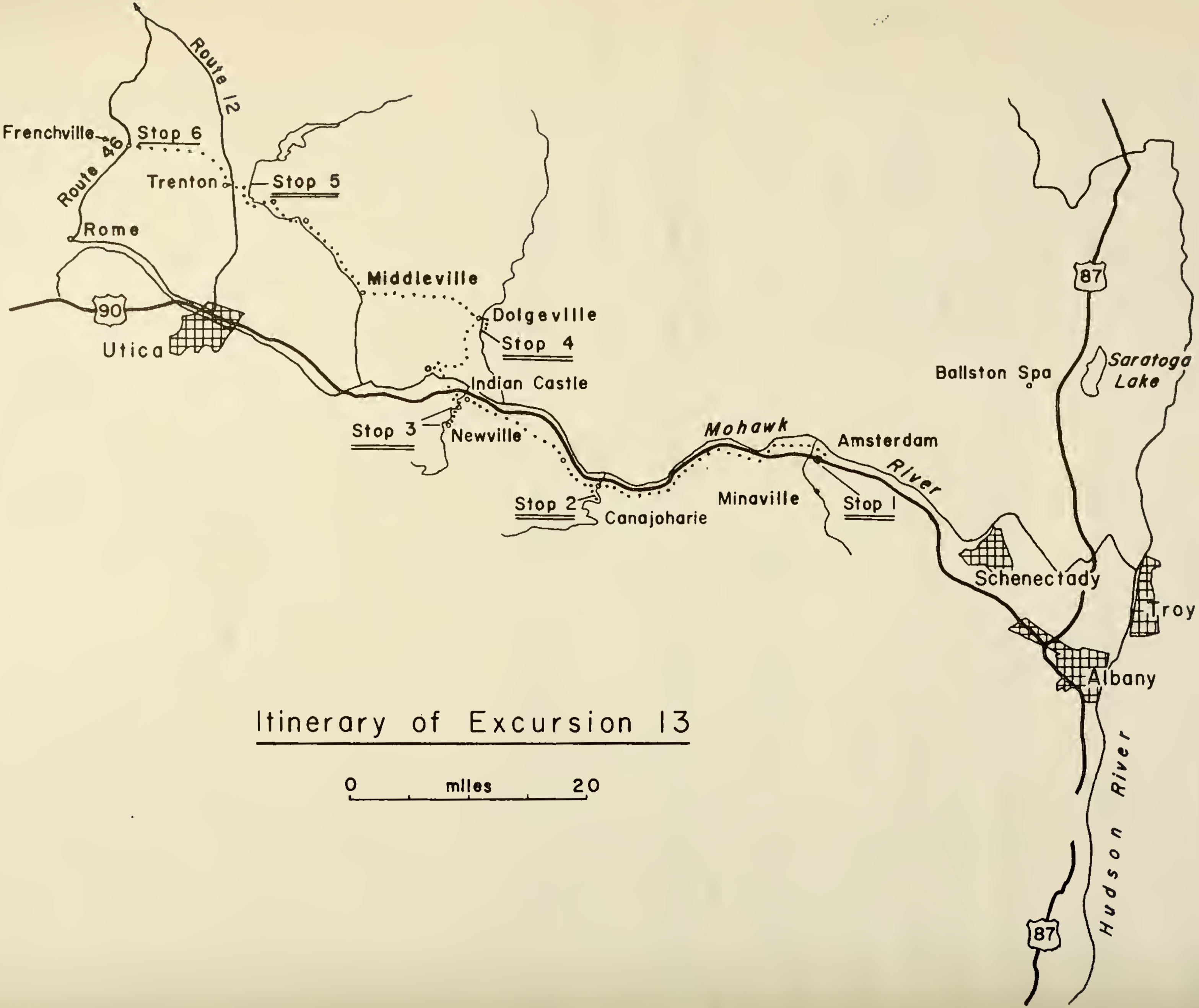

Itinerary of Excursion 13

Utica and Canajoharie Shales in the Mohawk Valley

Leader: John Riva

INTRODUCTION

In Middle Ordovician time carbonate deposition in the Appalachian miogeosyncline, from Pennslyvania to the island of Anticosti in the Gulf of St. Lawrence, was gradually replaced by an influx of black shale commonly referred to as the "Utica" shale. The "Utica" entered the miogeosyncline from the east or southeast and progressively moved westward attaining the cratonal margins in Ontario (Manitoulin Island) and well overlapping the Precambrian Shield (Lake St. John area in Quebec) in late mid-Ordovician time. This black shale invasion was followed by coarse clastic sediments, first represented by the Schenectady beds in the lower Mohawk valley, northwestern New York, and parts of Ontario and Quebec.

The "Utica" consists essentially of grey-black shale and thin limestone interbeds; lithologically, shales in the upper Mohawk valley are indistinguishable from those, much older, in the lower Mohawk. The combined thickness of the "Utica" is probably close to 2000 feet in the Mohawk valley and, possibly, as much as 3000 feet at the northern end of Lake Champlain in southern Quebec. The fauna of the Utica consists of brachiopods, trilobites (especially various forms of Triarthrus), various types of nautoloids, and abundant graptolites. The latter evolved rapidly throughout the "Utica" shales of all northeastern North America and afford us a sure measure of biostratigraphic, as well as stratigraphic, control which, otherwise, would by entirely lacking in a unit made up of black, brownish-weathering shale.

Ruedemann (1912, 1925a) divided the "Utica" into two broad biostratigraphic units: the Canahoharie Shale and the Utica proper, or the "true Utica". The Canajoharie comprises the older "Utica" shale in the lower Mohawk valley, where it can be recognized as far east as Ballston Spa and extends as far west as Dolgeville, East Canada creek, and Indian Castle. The Utica proper, as redefined by this writer (1968, 1969), includes most of the younger "Utica" shales in the upper Mohawk valley where they replace westward upper Trenton (Cobourg) limestone, and a thin sheet of uppermost Utica continues over the Trenton into the Black River valley, northwestern New York and Ontario. Basal Utica shale also occurs in the lower Mohawk valley (its most easterly outcrop is that along the Northway cut, 20 miles north of Albany) but the middle and upper Utica is replaced by the sandstone and shale of the Schenectady Formation. Thus the "Utica" is a classic time-transgressing unit which becomes progressively younger as one moves westward, to be superseded, in turn, by coarse clastics forming a number of units also time-transgressive.

The graptolite faunas of the "Utica" were only given a passing notice by early New York and Canadian geologists who believed them to be identical to those--now known to be much older-- of the Normanskill beds of the Hudson River valley, although Lapworth (1886), basing himself on the British graptolite succession wid was then in the process of erecting, thought otherwise. It remained to Ruedemar (1900) to show that Normanskill faunas had little in common with, and were i much older than, those of the "Utica". Subsequently, he recognized (1912, 1925a b) not less than nine graptolite zones in the Canajoharie-Utica succession of the Mohawk valley. Not all these zones, however, were based on clearly distinct fauna for some tended to accentuate the sudden appearence of new graptolites or the loca abundance of others; for these and other reasons Ruedemann's zones have not gained general acceptance in eastern North America.

The writer has for some time been conducting a painstakingly detailed study f the "Utica" graptolite succession both in the St. Lawrence Lowlands and in the Mohawk valley, and the results, although not yet in definite form, are shown graphically in Figure 1. "Utica" faunas in eastern Canada are identical to thce in New York. To the left of Figure 1 is Ruedemann's zonal succession; to the right a much simpler one proposed by the writer (1968, 1969) which preserves al valid Ruedemann zones, but proposes new ones to replace or amalgamate those whi n were not based on actual faunas. Two distinct faunas have been recognized in t e Canajoharie, and two, but strongly overlapping, faunas in the Utica proper. Th lower limit of the Utica has been lowered to coincide with the beginning of the Climacograptus spiniferus zone (Fig. 1).

The purpose of the excursion is to show some of the most typical outcrops o: Canajoharie-Utica shales and their characteristic faunas in the lower and uppe: Mohawk valley. The contact with the underlying or interdigitating Trenton lime stone will be studied in the most classical localities. A total of six (possily seven) stops are planned, beginning with the lower Canajoharie shale (and the Canajoharie-basal Trenton contact) opposite Amsterdam.

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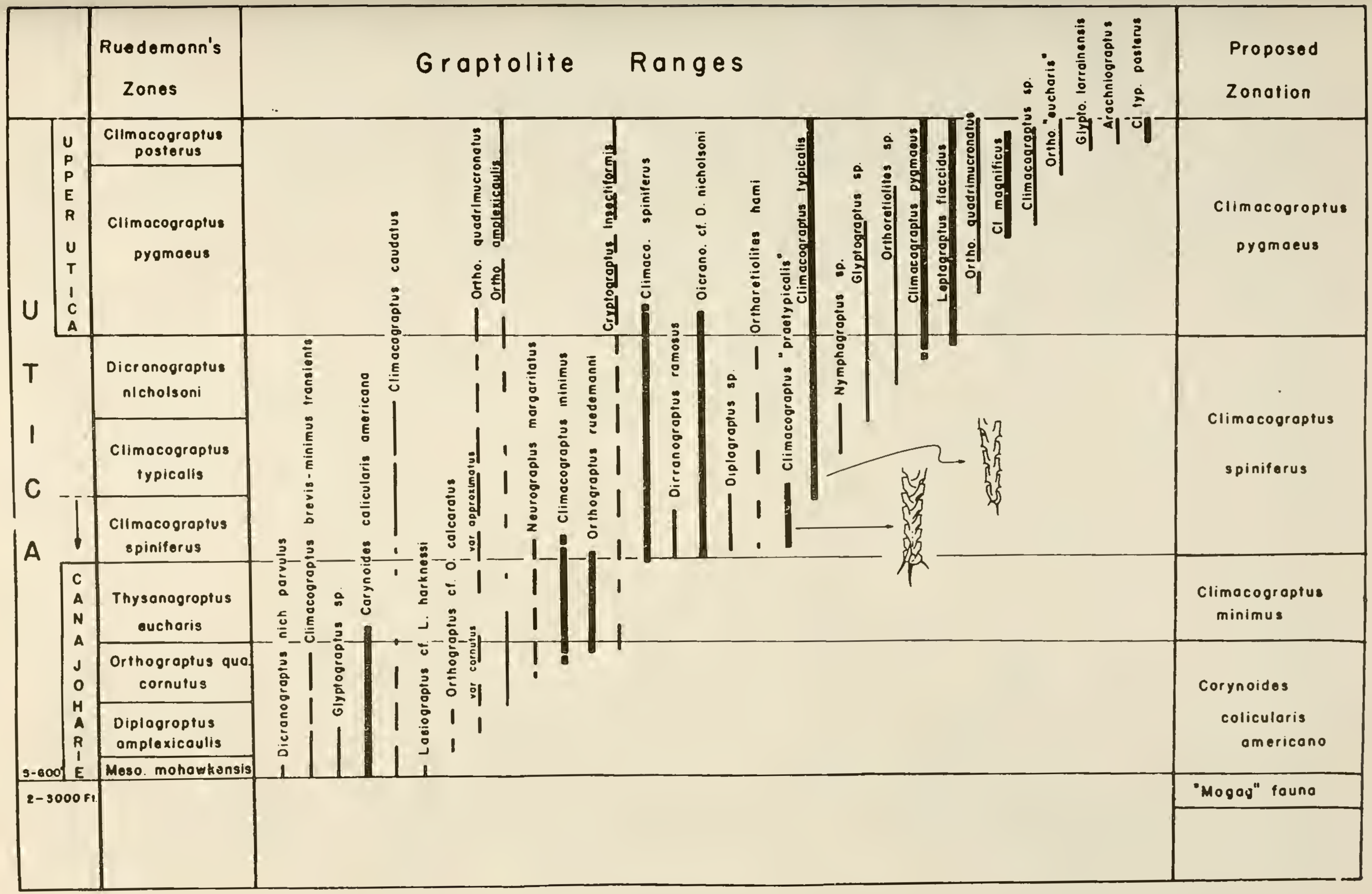

Fig.-1. Canajoharie— Utica graptolite faunas and their proposed zonation.

TRIP 13

Utica and Canajoharie Shales in the Mohawk Valley

Leader: J. F. Riva

LUNCH: Lunches will be provided for all those who have registered for the trip.

START: The party will depart from the Thruway Inn at Albany at 8:00 A.M. , Sunday, October 12, 1969, and proceed westward on the N.Y. State Thruway to the Amsterdam exit (Interchange 27). On leaving the Thruway, take N.Y. 5S and continue westward a short distance to the bridge on Chuctanunda creek. The outcrops on the Thruway between Interchanges 25 and 27 are of greenish-grey sandstone and shale belonging to the Schenectady Formation of the lower Mohawk valley. The type exposures of the Schenectady are along the banks of the Mohawk river east of Schenectady, where N.Y. 146 crosses the river between Aqueduct and Rexford. About seven miles east of Amsterdam, the Hoffman Ferry Fault brings up the Canajoharie and most of the Middle and Lower Ordovician section.

STOP 1: Bridge on N.Y. 5S over Chuctanunda creek. Park below bridge on the eas side. Exposed along the bed of the creek, on the walls of the gorge up to bridge level and along the road cut to the west, is the best section of Lower Canajoharie shale in eastern North America. The entire fauna of the zone of Corynoides calicularis var. americana is present, and the upper 20 feet also contain all the faunal elements of the next, and upper Canajoharie zone, that of Climacograptus minimus (Fig. 1). Corynoides calicularis occurs in great numbers throughout the section ar can be collected from handy outcrops at the corner of Daniel Street and Florida Avenue. In the lower beds of the creek, to the top of the lowe falls, a form of Glyptograptus is predominant. Orthograptus amplexicaulis, Orthograptus quadrimucronatus approximatus-cornutus, and the first of th "Utica" Triarthri, Triarthrus becki, first appear in a cut west of the bridge. The basal part of the Canajoharie is covered, as is the contact with the underlying Trenton limestone, Here, judging from Prosser, Cummins, and Fishers' map (1900, N.Y. State Museum Bull. 34), about 20 to 30 feet of basal shale are missing. The Trenton (Shoreham Limestone)—Canajoharie contact is well exposed, however, three miles to the east on Morphy's creek, a locality which can now be reached only by traveling east on the Thruway.

Outcrops fail for some distance above the bridge, but reappear in the bed of the creek just below and above Minaville. Here graptolites of the second zone of the Canajoharie are the only fossils present.

From Chuctanunda creek proceed to Canajoharie following N.Y. 5S. Canajoharie shales crop out on the left side of the road to Fultonville. Just past Randall, a small exposure of Precambrian gneiss occurs on the left side of the road. From here to Canajoharie the cliffs on both sides of the Mohawk valley consist of Lower Ordovician and Upper Cambrian dolomites and minor limestone. The Ordovician-Cambrian contact is exposed in an abandoned quarry on the left side of the road as one reaches Canajoharie. On entering the town, turn left at the first traffic light, drive across the railroad tracks and take the narrow road on the west side of the creek to the very end. (An excellent field guide for the Lower Ordovician of the Mohawk valley has been prepared by Dr. D.W. Fisher of the N.Y. State Museum, Educational Leaflet 18, 1965.)

STOP 2: Canajoharie shale at Canajoharie creek. Type section of the Canajoharie. At the lower falls are massive Lower Ordovician dolomites and limestone, with very few fossils, followed disconformably at about dam-level by 17 feet of lower Trenton (Shoreham) limestone and shaly interbeds. Black River limestones are absent between the two units. The Shoreham Limestone bears abundant trilobites, brachiopods, and bryozoans; the shaly interbeds have graptolites. Canajoharie shale abruptly follows the Shoreham and the contact—visible at foot level around the bend of the creek above the dam—is marked by an uneven, rusty surface which in the "Utica" always denotes a hiatus, the extent of which can usually be determined by strict faunal control. Here it appears that part of the lower Canajoharie at Chuctanunda is missing. The type Canajoharie is about 250 feet thick and contains graptolites of the upper part of the zone of <u>Corynoides calicularis americana</u> and those of the zone of <u>Climacograptus minimus</u>. The true Utica does not appear at the top of the section, making it incomplete both at the top and at the base.

Return to Canajoharie and continue westward on N.Y. 5S past Fort Plain to the hamlet of Indian Castle and Nowadaga creek. Take the paved Creek Road on the west side and ascend the creek to the first bridge. Stop and park on either side of the bridge.

STOP 3: Utica shale proper. Well exposed in Nowadaga creek, from the first bridge to, and above Newville, and then along Ohisa creek (a tribuitary of Nowadaga creek).is the most continuous, although incomplete, section of true Utica in the Mohawk valley. The section is closely matched by a similar one along the northern shore of the St. Lawrence river at Neuville, which, however, is accessible only at low tide. The contact with the underlying middle Trenton limestone is covered, but isolated outcrops of transitional limestone-shale beds (usually referred to as the Dolgeville facies) are found on the right bank of the creek below the first bridge. The actual Trenton-Utica transition will be examined at the next stop at Dolgeville; an excellent, although inaccessable, outcrop showing this transition is exposed in a cut on the N.Y. Thruway just west of the Indian Castle rest stop, a short distance west of Nowadaga creek.

The Utica exposures along the creek are about 600 feet thick. The graptolite fauna from the first bridge to just below the village of Newville,belongs to the zone of <u>Climacograptus spiniferus</u>. Basal Utica graptolites, including the predecessor to <u>Climacograptus typicalis</u>, occur between the first and second bridge. <u>Dicranograptus</u> cf. <u>D. nicholsoni</u>, <u>Orthograptus quadrimucronatus approximatus</u>, <u>Climacograptus spiniferus</u>, <u>C. typicalis</u>, <u>Thriarthrus eatoni</u>, straight nautoloids, ostracods, and various brachiopods occur through the exposure to Newville. At Newville, at and above the bridge, there is a distinct change in fauna, and graptolites are those of the upper Utica zone of Climacograptus pygmaeus. Common here is a form of Glyptograptus, a short form of Corynoides (not yet identified in the St. Lawrence Lowlands), C. typicalis, and Leptograptus flaccidus.

From Newville drive back to N.Y. 5S and proceed west, taking the turn-off to Little Falls. At the first traffic light at the east edge of town, turn east on N.Y. 167 to Dolgeville. Go through Dolgeville and cross East Canada creek at the first available bridge. Continue south on the road on the east side of the creek for about one mile. Park, and take a footpath to the creek.

STOP 4: Dolgeville beds and transition into the true Utica. Exposed below the falls, on the east side of East Canada creek, are what have long been referred to as the Dolgeville beds, or facies, which continue westward for more than 25 miles to West Canada creek and show the interfingering of the "Utica" with the Trenton limestone. The Dolgeville beds have been correlated with middle Trenton Denmark limestone of Trenton Falls. Lithologically, they consist of platy limestone alternating with shale beds of equal thickness. Graptolites from the shales, although few in number, are those of the upper Canajoharie zone of Climacograptus minimus. The Canajoharie shale apparently extends no farther west than East Canada creek and Nowadaga creek, a few miles to the south. Above the Dolgeville beds, the shale at the falls, above the dam, and through Dolgeville, contains lower Utica graptolites very much like those in the beds between the first and second bridge on Nowadaga creek.

Return to Dolgeville, take the upper bridge and drive northwest on the Military Road to Salisbury. Turn west on Route 29 to Middleville. Here take route 28N and pass through Newport, Poland, and Gravesville to West Canada creek. Shortly after the bridge over West Canada creek take the turn-off (not to be missed) to the hamlet of Trenton Falls. Go through the hamlet into the property of Niagara Mohawk Power and park at the fountain. Continue on foot to the highest tower and the pipelines. Walk west on the pipelines (if provided with rubber shoes, and at your own risk) until the High falls of Trenton Falls come into view, below.

STOP 5: Type section of the Trenton at Trenton Falls. The middle Trenton (Denmark) limestone appears just above the base of the upper High falls, comprises the lower falls, and continues eastward down to, and through, Sherman falls, out of view to the right. The beds just below Sherman falls contain many three-dimensional Orthograpti amplexicaules, one of the most common graptolites of the upper part of the lower graptolite zone of the Canajoharie, the zone of Corynoides calicularis americana (already seen on Chuctanunda creek). The basal part of the Trenton, corresponding to the Shoreham Limestone, is not exposed at Trenton Falls. The uppermost seven feet of Denmark limestone begin at the platform above the lower falls and include the basal portion of the upper High falls. They consist of alternating platy limestone and shaly interbeds, similiar to those at Dolgeville, and could be considered the most western extension of these beds. No graptolites have yet been collected from these beds, although trilobites, brachiopods and other fossils abound; the proposed correlation with the beds at Dolgeville is based purely on lithologic similarity.

Most of the upper High falls, the beds in the creek all the way up to dam, those at the spillway, and those forming the walls of the gorge are of upper Trenton (Cobourg) limestone. This limestone is highly fossiliferous and contains a characteristic brachiopod, Rafinesquina deltoidea, which is widely distributed in the Cobourg of northeastern North America. At the next stop, eleven miles west as the crow flies, the Cobourg beds will be, in turn, overlain by uppermost Utica shale, albeit disconformably. (For comprehensive and detailed studies of the Trenton of the Mohawk valley, consult the following papers by Marshall Kay: a. Geology of the Utica Quadrangle, New York: N.Y. State Museum Bull. 347 1953; and b. Stratigraphy of the Trenton Group: Geol. Soc. America Bull., vol. 53, p. 233-302, 1937.)

Drive back to Trenton Falls and turn right. When the road splits take the left branch, continue on through an underpass, and across U.S. 12 to the village of Trenton (Barneveld). At Trenton turn right and take county Route 274 towards Frenchville. About two miles east of Frenchville (and the intersection with Route 43), the road descends into a low gorge, the walls of which are formed of brown-black Utica shale. Exposed in the bed of the creek, locally known as Big brook, is Cobourg limestone. Stop and park at a handy turn-off on the left side of the road (proceeding west). Walk across the creek bed to the Utica-Trenton contact.

STOP 6: Uppermost Utica shale on Cobourg limestone. An excellent contact is on the left bank of Big brook. The Cobourg limestone is in massive beds with no trace of shaly interbeds. The Utica shale that abruptly follows the Cobourg rests, as at Canajoharie and most other localities in northwestern New York, Ontario, and the upper St. Lawrence Lowlands, on a surface of disconformity, the vertical extent of which can be determined only where graptolites are present above and below it. The surface of the upper Cobourg bed is covered in part by a rusty layer caused by the weathering of a thin pyrite layer at the base of the Utica shale. In the writer's experience, pyritic layers within the "Utica" or at the Trenton-Utica contact indicate a hiatus of variable length and extent.

Graptolites in the black shale are those of the upper Utica zone of Climacograptus pygmaeus. Interbedded with the black shale are dark grey, micaceous shale and siltstone typical of the basal beds of the onlapping Lorraine beds.

End of the excursion. Participants returning to Albany should take Route 46S to Rome, Utica and the Thruway, or return to Trenton and take Route 12S to Utica and the Thruway. Participants heading for Canada should take Route 46N to Boonville, Route 12 to Watertown, and then Interstate 81 to the Thousand Islands Bridge and Ontario.

Trip 14

PALEOECOLOGY AND STRATIGRAPHY OF ORDOVICIAN CARBONATES, MOHAWK VALLEY, NEW YORK

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INTRODUCTION

The Ordovician of the Mohawk Valley (Fig. 1) offers a rather good opportunity for the study of shallow-water carbonate biofacies. The regional stratigraphy suggests that this was an area of more or less predictable environmental variations. The carbonate units are contemporaneous with black shale facies to the east: furthermore, the area is bisected by the Adirondack Arch (Fig. 2), a positive feature during early Paleozoic time. However, the lateral changes in lithologic facies and the discontinuous distribution of the units impose some difficulty in the determination of time equivalencies. Therefore, it is difficult to arrive at any sort of meaningful paleogeographic synthesis; each section has to be interpreted on its own merit, and regional comparisons of paleoenvironments must be very generalized.

The following discussion of the paleoecology of the area is meant to serve as an example of the quantitative approach. Several analytical techniques are used to derive classifications and delineate trends for the biofacies. For those of you not familiar with these techniques reference is made to key papers to supplement the necessarily brief explanations given here. For all trip participants there will be ample opportunity to compare the quantitative abstractions with the "real world" as represented at the various stops.

The Middle Ordovician (Mohawkian) fossils can be classified into biofacies, representing groups of fossils that tend to occur together, and biotopes, representing groups of samples that contain similar types of fossils. Both types of classifications can be effected readily by using cluster analysis (Parks, 1966; Harbaugh and Merriam, 1968). First, samples are obtained; generally our procedure has been to count, or note the presence or absence of, various major types of fossils in a bedding-plane sample area of 600 cm^2 (roughly the area of a sheet of paper). Then, to obtain a biotope classification,

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1/ Park is primarily responsible for the discussion of the paleoecology and Fisher is primarily responsible for the stratigraphic interpretations. Much of the faunal data was collected by Leonard A. Porter as part of research sponsored by an NSF Teacher Research Participation grant.

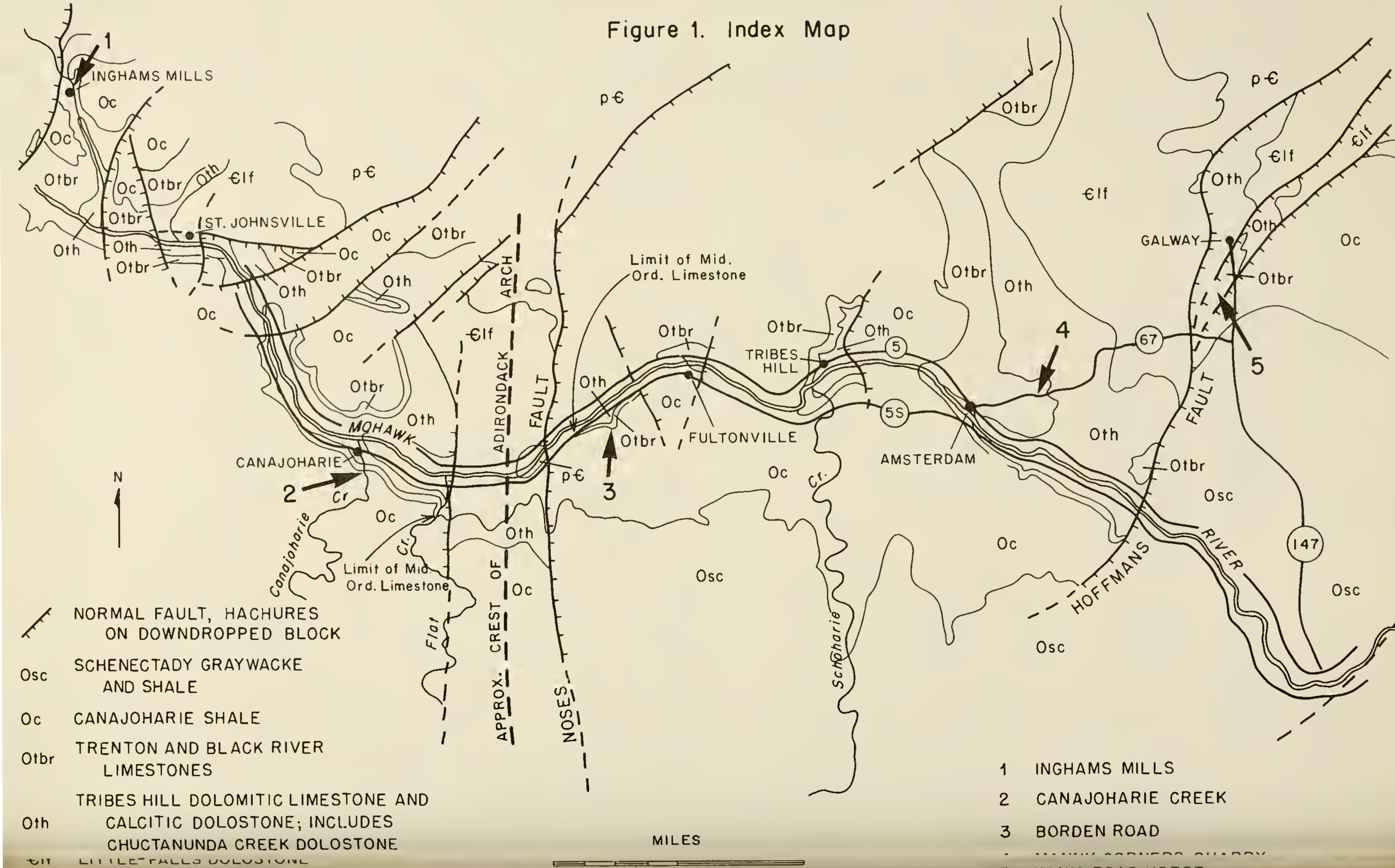
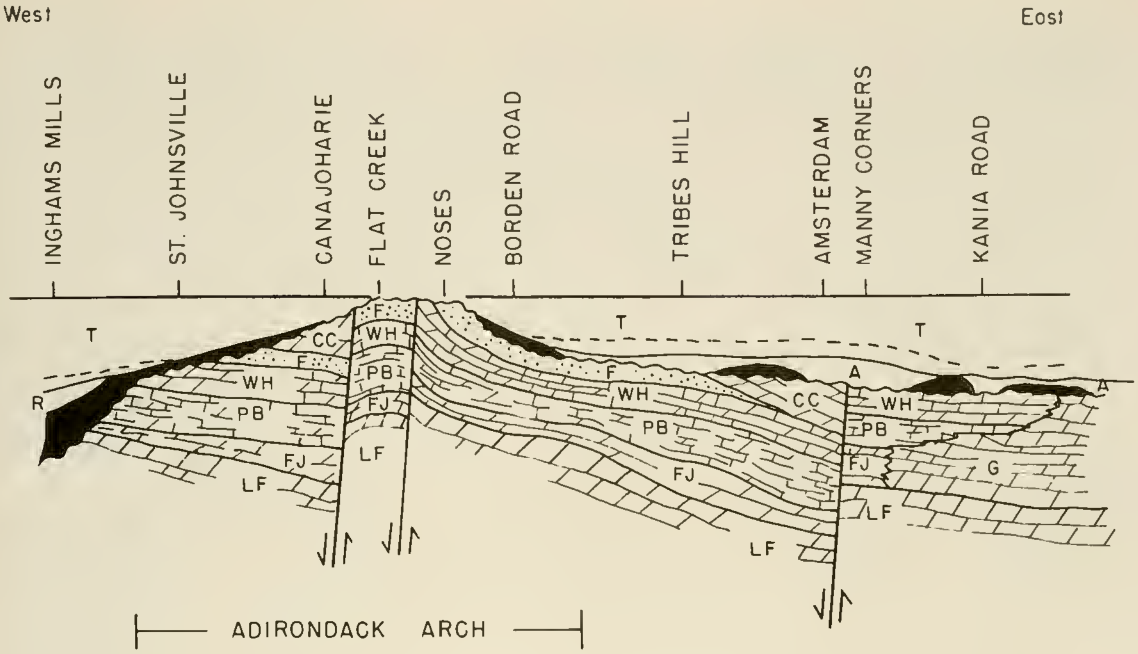


Figure 1. Index Map

Figure 2. Stratigraphic Cross-section

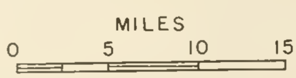


Datum is base of Canajoharie Shale

T	TRENTON LIMESTONE (includes Sugar River and Kings Falls Limestones west of the Adirandock Arch and Glens Falls Limestone east of Arch)
A	AMSTERDAM LIMESTONE
R	ROCKLAND LIMESTONE at Inghams Mills
L	LOWVILLE LIMESTONE
CC	CHUCTANUNDA CREEK DOLOSTONE
F	FONDA LIMESTONE
WH	WOLF HOLLOW LIMESTONE
PB	PALATINE BRIDGE LIMESTONE
FJ	FORT JOHNSON DOLOSTONE
LF	LITTLE FALLS DOLOSTONE
G	GAILOR DOLOSTONE

TRIBES HILL FORMATION

NOTE: Post-Lower Ordavician (Canadian) faults not shown



each sample is quantitatively compared with every other sample on the basis of similarities in numbers of the various faunal types; this can be done using correlation coefficients or any one of a number of other similarity measures. I prefer to use Sorensen's coefficient $2\underline{w}/(\underline{a}+\underline{b})$ where $\underline{w}$ is the sum of the lesser values for each of the variables (fossil types) in the two samples being compared, $\underline{a}$ is the sum of the values for all the variables in one sample and $\underline{b}$ is the sum of the values for all the variables in the other sample. A dendrogram is then constructed by clustering the two samples with the highest similarity, averaging them, recalculating their joint similarities with all other samples, clustering the pair of samples, or clusters, having the next highest similarity, etc. The samples are represented by the horizontal lines and their levels of similarity are indicated by the position of the vertical lines of juncture. The result is an objective hierarchical classification of all the samples based on however many equally-weighted variables the investigator wishes to consider. A classification of biofacies may be similarly derived by comparing fossil groups on the basis of their joint abundances in the samples.

Several biofacies can be distinguished within the Glens Falls and Amsterdam Formations (Fig. 3). These are recognizable in the field, but they are not well defined, as indicated by the low levels of clustering. Therefore, any one of the biofacies is likely to occur with some degree of admixture of representatives from one or more other biofacies. The very low levels of clustering of some of the fossil types may be a function of time, the fossils not having evolved until part way through the time period represented, or of rarity, the fossils occurring so infrequently that their true ecologic affinities are not adequately represented. It is interesting to note that each of the biofacies is composed of a single taxon except for the coral biofacies, which includes solitary and colonial corals and the alga Solenopora. The brachiopod biofacies can be subdivided into two facies of slightly different depositional history; 1) a facies of flat-lying, concave-down brachiopods which probably accumulated in tidal flats and very shallow subtidal areas and 2) a facies of imbricate, concave-up brachiopods which probably represent a lag deposit of migrating tidal channels. All the biofacies are thought to represent normal-salinity intertidal and shallow subtidal environments. They differ in their adaptation to turbidity, wave energy, substrate firmness, and possibly water depth (including the effects of dessication and light penetration).

The biotopes of the Glens Falls and Amsterdam Formations are more discernible (Fig. 4), comprising two main clusters of samples which can be interpreted as representing muddy and clear-water environments. The "clear-water" biotope consists primarily of brachiopod, echinoderm, and gastropod biofacies; the "muddy" biotope consists of trilobite and ectoproct (bryozoa) biofacies. The environmental interpretations are based on a consideration of the functional morphology of the fossils and, to a lesser extent, on their lithologic associations. In general the Amsterdam and lower Glens Falls (Larrabee Limestone) represent dominantly "clear-water" high-energy, subtidal to intertidal biotopes and the upper Glens Falls (Sugar River Limestone) equivalent represents dominantly "muddy", lower-energy, subtidal and intertidal biotopes but with intercalated "clear-water" beds.

Lateral faunal trends from one locality to the next are difficult to detect, probably because of the poor stratigraphic control and the slight differences in the environmental responses of the faunas. It is likely that there were no consistent environmental gradients (for instance, no sharp depth trend) in the study area during the Middle Ordovician; therefore, the faunas did not vary consistently, but fluctuated depending on local conditions. However, this does not mean that regional differences, such as might be predicted by the regional stratigraphy, did not exist. If the average faunal compositions of the Glens Falls equivalents west of the Adirondack Arch are compared with those for the Glens Falls Formation east of the Arch (Fig. 5) it can be seen that trilobites and ectoprocts are more common to the east and brachiopods and echinoderms are more common to the west. It is reasonable to infer that the organisms were responding primarily to the amount of fine-grained terriginous clastics, expressed as turbidity and substrate softness; therefore, this regional difference is probably related to the influx of clastics from the newly emerging Taconics to the east. The difference appears to have been more pronounced during the latter part of the time span (the Shorehamian Substage), as one might predict.

It is instructive to perform time-trend analysis (Harbaugh and Merriam, 1968) on the sequences of biofacies in individual stratigraphic sections in order to test for significant vertical trends at the local level. The section at Inghams Mills (Stop 1) is given as an example. Time-trend analysis is essentially a bivariate technique; that is, it is concerned with the variation of only one variable plotted against time, the independent variable. However, it can be applied to multivariate data, consisting of a number of variables such as fossil types, by assigning numerical values to the variables so as to place them in a logical progression. At Inghams Mills each bed in the upper part of the section (the Glens Falls Formation correlatives, the Kings Falls and Sugar River Formations) was identified as to biofacies using the classification derived from cluster analysis. The biofacies were numbered in order of increasing tolerance to mud, and a point, representing the scaled biofacies, was plotted for each bed. The points were then smoothed (Fig. 6) and were tested for significant trends (deviation from vertical), random fluctuations, and cyclicity. The resulting curve probably does not indicate anything more than a careful investigator would note in the course of examining the section. However, by quantifying the study it is possible to determine <u>in an unbiased manner</u> that: 1) there is a definite faunal trend with respect to our inferred ecologic progression; 2) the fluctuations in dominant biofacies are probably not random; and 3) there is no pronounced cyclicity. Time-trend analysis of the upper part of this local section indicates a gradually changing environment, probably brought about by the increasing effects of fine-grained terrigenous clastics from the east.

In summary, quantitative paleoecologic analysis, as shown in these few examples, seems to be a powerful tool in determining paleoenvironments. The actual comparison between these quantitative models and the "real world" is left up to you, the field trip participant!

STOP 1, INGHAMS MILLS

Location of section (Figs. 1, 2): Little Falls 7.5' quadrangle. Rock exposed in East Canada Creek just north of village of Ingham Mills, a hamlet 5.7 miles northwest of St. Johnsville and 4.7 miles east-northeast of Little Falls. Section on property of Niagara-Mohawk Power Corp. (Prosser, 1900; Kay, 1937; added info. by Park and Fisher)

Discussion: (Note: This section will also be visited by Trip 16 on Sunday, at which time its stratigraphic implications will be discussed more thoroughly, especially with respect to the Black River Valley outcrop belt.) At very low water buff weathering dolostone has been reported. Ordinarily, the oldest strata visible are massive white weathering, brownish-gray ("dove") lithographic limestones with a pronounced conchoidal fracture. These lime muds represent ancient intertidal flats penetrated by vertical worm burrows (Phytopsis) and exhibiting filled dessication cracks. Occasionally, interbeds of lumpy shaly limestone occur. Some beds weather buff reflecting a dolomitic component. Halfway up the steep wall, a darker gray weathering, medium-dark gray, non-conchoidal limestone occurs. This contains subtidal fossils (brachiopods, gastropods, horn corals) and is interpreted as representing an interfingering of the marine Chaumont facies. The overlying strata exhibit a return to intertidal Lowville conditions with tidal channels, Phytopsis, and the radially branching burrow Chondrites; a bed of Tedradium corals and ostracodes lies 2 feet below the prominent bench.

Conformably on the brecciated uppermost Lowville bed is 12 feet of brownish-weathering nodular black argilli-calcilutite with shale interbeds. This is the Rockland Limestone. It contains abundant detritus-feeding trilobites, especially Isotelus, and occasional layers of brachiopods, especially Doleroides in the lower part. The upper part of the unit contains a more diversified fauna of gastropods, cephalopods, brachiopods, and fodinichnia (trace fossils made by burrowing detritus feeders); this is primarily a "clear-water" assemblage in contrast to the "muddy" assemblage of the lower Rockland.

Above the Rockland Limestone is about 4 feet of gray calcarenite with abundant clasts of Lowville Limestone and rare dolostone clasts (Little Falls Dolostone?). This suggests a moderate relief within close proximity to the area of deposition. This limestone is a thin representation of the unit newly named Kings Falls Limestone by Kay (1968). The basal bed contains abundant gastropods as well as clasts and is interpreted as denoting a very shallow water, hard substrate environment (rocky intertidal?). The succeeding beds include 2 rippled horizons and, for the most part, have been completely reworked by burrowing organisms.

The Sugar River Limestone (Kay, 1968) follows to the dam. This consists of interbedded calcarenites and calcilutites, with the argillaceous content appearing to increase upward. The fauna is quite diverse. The lower part of the unit contains concave-down and imbricate, concave-up brachiopod biofacies (Fig. 3); the trilobite biofacies is more common in the middle part of the unit; and the upper part is dominated by the ectoproct (bryozoan) biofacies. The strophomenid brachiopods are interpreted as being somewhat similar to oysters in mode of life; the imbricate brachiopod biofacies probably represents the lag concentrate of tidal channels that slowly migrated across the Middle Ordovician tidal flats. The thin, fine-grained interbeds, with Flexicalymene and other members of the trilobite biofacies, were probably tidal flat deposits that escaped reworking. The ectoproct biofacies is considered to be a subtidal, "muddy" assemblage. This is in contrast to the Cenozoic ectoprocts which are intolerant of mud. See the INTRODUCTION for a further discussion of the time-trend analysis of this part of the section.

STOP 2, CANAJOHARIE CREEK

Location of section (Figs. 1, 2): Canajoharie 7.5' quadrangle. Rock exposed along Canajoharie Creek (public) at southern edge of the village of Canajoharie. Reach section from highway N.Y. 5-S by taking Moyer St. south (uphill) and turning right (west) on Floral Ave. to its end. Park cars on village property between greenhouse and gorge. (Kay, 1937; Fisher, 1954)

Discussion: The lowest strata exposed are intertidal, supratidal, and shallow subtidal facies of the Tribes Hill Formation and are about 0.3 mile downstream from the descent ladder. The Wolf Hollow Member (dolomitic calcilutite) is conformable on the uppermost beds of the Palatine Bridge Member (thin silty limestones and shale interbeds). The Fonda Member of the Tribes Hill Formation overlies the Wolf Hollow Member and is just around the bend in the creek. The large pothole which gives Canajoharie its name (Canajoharie is the Iroquois name for the "pot that washes itself") together with several other smaller potholes, is in the Chuctanunda Creek Dolostone, the youngest Early Ordovician unit in the Mohawk Valley. Aside from the "hippopotami backs", which are dolomitized hemispherical stromatolites (algal mounds), no fossils are known from the Chuctanunda Creek Dolostone. The environment of deposition is probably lower intertidal.

Disconformably resting on the Lower Ordovician carbonates is the Middle Ordovician Trenton Limestone (Sugar River?). Here, this limestone consists of a few calcarenite beds with nodular and irregularly bedded argilli-calcilutites with shale interbeds. The limestones contain brachiopod, ectoproct, and trilobite biofacies. Only 15 feet of Trenton limestone occurs here and 4 miles to the east, the Trenton is absent and the Canajoharie Shale rests directly on the Chuctanunda Creek Dolostone. Here, the Trenton is on the western flank of the Adirondack Arch, whose crest lies slightly to the east along the longitude of the Noses Fault midway between Canajoharie and Fonda.

The Canajoharie black shale can be seen high on the gorge walls with an abrupt but conformable relationship to the Trenton Limestone. Upstream at Wintergreen Park is a spectacular gorge exposing some 200 feet of Canajoharie black shale----its type locality. We will eat lunch at Wintergreen Park.

STOP 3, BORDEN ROAD

Location of section (Figs. 1,2): Randall 7.5' quadrangle. Borden Road leaves N. Y. 5-S, 2.7 miles west-southwest of Fultonville and slightly east of where Van Wie Creek passes under N. Y. 5-S. The section begins in Van Wie Creek (private) and continues along the east side of Borden Road (public property) and re-appears southward along Van Wie Creek (on posted property!) (Fisher, 1954)

Discussion: The Palatine Bridge and Wolf Hollow Members of the Tribes Hill Formation outcrop in the creek. In the ditch on the east side of Borden Road is a fossiliferous light gray glauconitic calcarenite and pebble conglomerate, characteristic of the Fonda Member. It is replete with the pelecypod-like arthropod known as *Ribeiria*. The trilobites *Clelandia*, *Hystricurus* and *Symphysurus* occur as do many species of snails (*Ophileta*, *Ecculiomphalus*, *Gasconadia*), brachiopods (*Finkelnburgia*, *Tetralobula*, a lingulid), nautiloids (*Ellsmereoceras*, *Ectenoceras*, *Clarkoceras*), conodonts, and probably cystid plates. The fauna is assuredly Early Canadian and is suggestive of a restricted marine environment, possible lagoonal.

A few more feet of intertidal dolomitic limestone and calcitic dolostone complete the Lower Ordovician section. Disconformably atop this karst surface are a few inches of "*phytopsis*"-bearing Lowville Limestone filling the widened joints in the Lower Ordovician dolostone. When this cut was fresh, brown wads of limonite were to be seen along this erosional surface. Middle Ordovician Trenton Limestone containing shallow-, clear-water biofacies (including the echinoderm biofacies) follows.

This section is significant in that it demonstrates conclusively the extent of the gap in sedimentary record (hiatus) present on the eastern flank of the Adirondack Arch. Of the Early Ordovician, the Middle and Late Canadian are absent. The Chazyan Series is wholly missing. Most of the Black River Group is missing. The early Trenton (Rockland) is absent. Certainly, the Adirondack Arch was an effective barrier to sedimentation and faunal dispersal during much of the Ordovician Period. There is some evidence that it was active, also, during the deposition of the Late Cambrian Little Falls Dolostone. However, there is no evidence that the Arch was effective in restricting sediments or faunas following deposition of the Middle Ordovician Schenectady graywackes and shales.

STOP 4, MANNY CORNERS QUARRY

Location of section (Figs. 1, 2): Amsterdam 7.5' quadrangle. Abandoned shallow quarry on south side of highway N. Y. 67, 0.3 mile east of Manny Corners, a hamlet 2.7 miles east-northeast of the main intersection in Amsterdam. Property of the Cushing Stone Corp. (Kay, 1937; added info. by Park and Fisher)

Discussion: The floor of the quarry is the Wolf Hollow Member of the Tribes Hill Formation; the Fonda Member and the Chuctanunda Creek Dolostone have been eroded. Numerous dolomitized cross-sections of the gastropod Ecculiomphalus may be seen. The fact that the fauna consists only of this one genus of gastropod, which has been diagenetically altered, suggests that this is an intertidal deposit. This gastropod bed near the top of the Wolf Hollow Member is relatively widespread in this area of the Mohawk Valley. The bed is overlain by 2 more feet of Wolf Hollow followed by 5 feet of light gray slightly quartzose calcisiltite with a slightly conglomeratic base. Kay (1937) referred to this unit as the Lowville but it exhibits neither the characteristic Lowville lithology nor the Lowville intertidal fauna. Brachiopods, gastropods and even an occasional horn coral may be found. This unit appears to be a representative of a localized environment along the southern flank of the Adirondacks during the period of Black River-Early Trenton deposition.

Twelve feet of Amsterdam Limestone follows. This consists of nodular to irregularly bedded dark gray to black fine- to medium-textured limestone with a prolific and varied fauna. Gastropods, horn corals, and brachiopods are particularly abundant, followed by trilobites, nautiloids, ostracodes, ectoprocts, and favositid corals. Pelecypods, conodonts, and scolecodonts complete the known fauna. The Amsterdam Limestone is restricted to a belt extending from Tribes Hill on the west to Glens Falls on the east.

Abruptly overlying the Amsterdam is the Larrabee Member of the Glens Falls Limestone. The Larrabee is a coarse- to medium-textured light gray regular bedded limestone, thicker bedded than any other of the Trenton limestones. Brachiopods are particularly abundant, exceeding all other fossil types. Echinoderm plates, trilobites and ectoproct are a poor second. The unit is dominantly a "clear-water", high-energy facies. The brachiopod Parastrophina hemiplicata and the trilobite Encrinurus cybeliformis appear to be reliable guides to this unit.

Along the north side of the highway is about 10 feet of tan weathering argilli-calcilutite with shale interbeds. This is the Shorehamian substage within the Glens Falls Limestone. Trilobites and ectoprocts are the dominant types. The trilobites Cryptolithus tesselatus and Isotelus gigas are particularly common here. The "chocolate-drop" ectoproct Prasopora is also easily found. The fauna is dominantly a "muddy" assemblage.

STOP 5, KANIA ROAD HORSE

Location of section (Figs. 1, 2): Pattersonville 7.5' quadrangle. Kania Road is the first east-west road south of, and parallel to the West Galway Road and is reached from Highway N. Y. 67 by taking Jersey Hill Road north 1.3 miles and turning right (east) on Kania Road. Section begins where road descends over Upper Cambrian Little Falls Dolostone and the Hoffmans Fault. (Public property) (Fisher, 1966)

Discussion: Between two major gravity faults, branches of the great Hoffmans Fault, downthrown on the east, is a sliver or horse of Middle Ordovician limestone itself bisected by an easterly down-dropped gravity fault. The western portion of the horse displays gently westward dipping Lowville Limestone conformably overlain by Larrabee calcarenite; the Amsterdam Limestone is absent from its accustomed position between the two! The eastern portion of the horse shows steeply eastward dipping light gray coarse calcarenite with golf ball-size pebbles of white to light graw weathering black limestone and algal nodules (*Solenopora*). It would seem that the Amsterdam has been eroded and redeposited as clasts within the Larrabee lime sand. The coral biofacies, including the abundant *Solenopora*, is thought to represent a very shallow water, high energy environment. The echinoderm biofacies, which also occurs here, interpreted as having been deposited in a lower energy environment at least slightly offshore. Careful search will uncover unusually fine specimens of rafinesquinid and sowerbyellid brachiopods. The contrast in stratigraphy and paleoecology between this locality and Stop 4; 6 miles to the southwest, is quite striking. Canajoharie black shale occurs on the eastern downthown side of the Hoffmans Fault.

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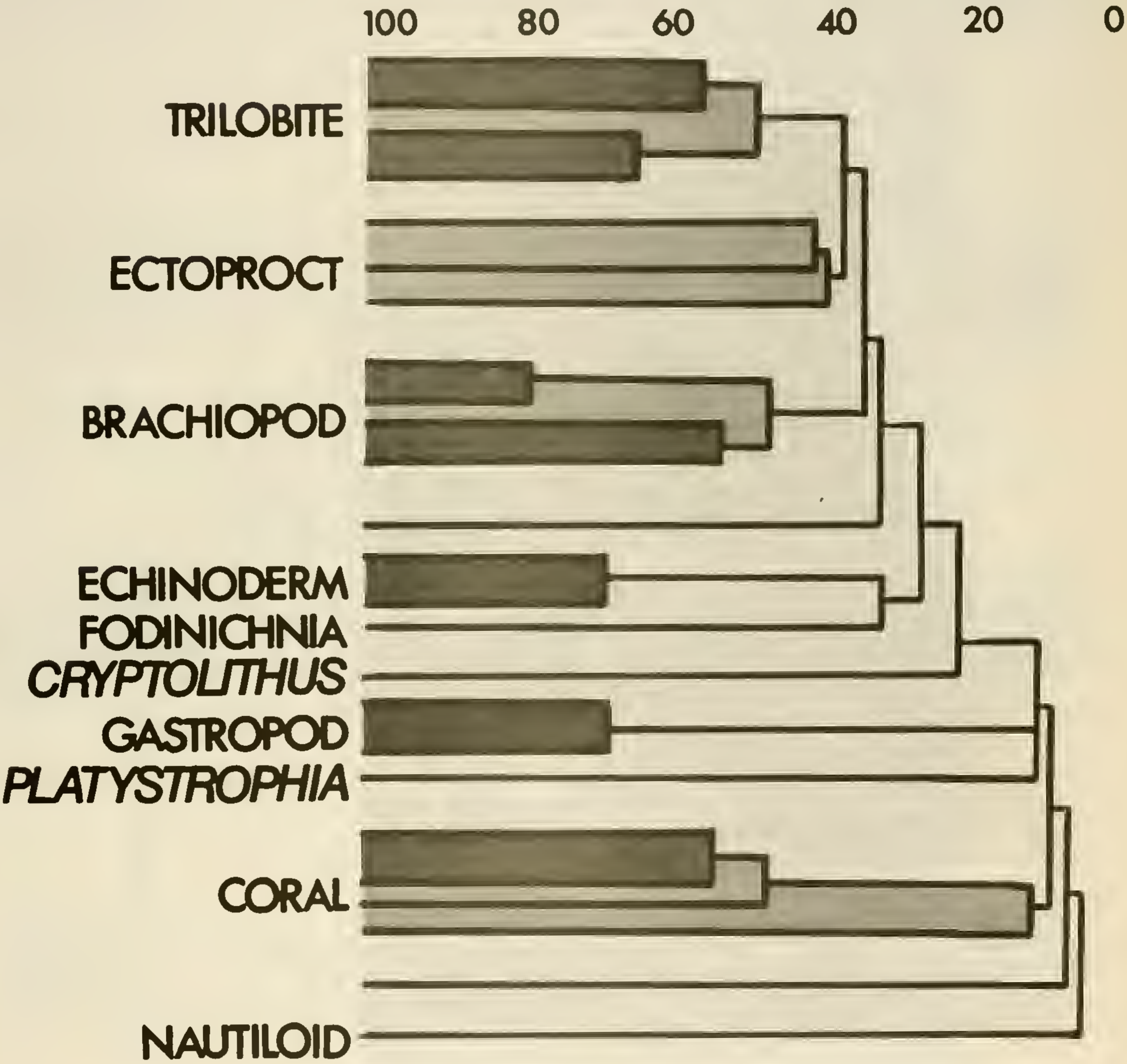

Figure 3. Biofacies classification based on cluster analysis; each horizontal line represents a fossil type; numbers refer to the similarity indices and indicate the levels of clustering.

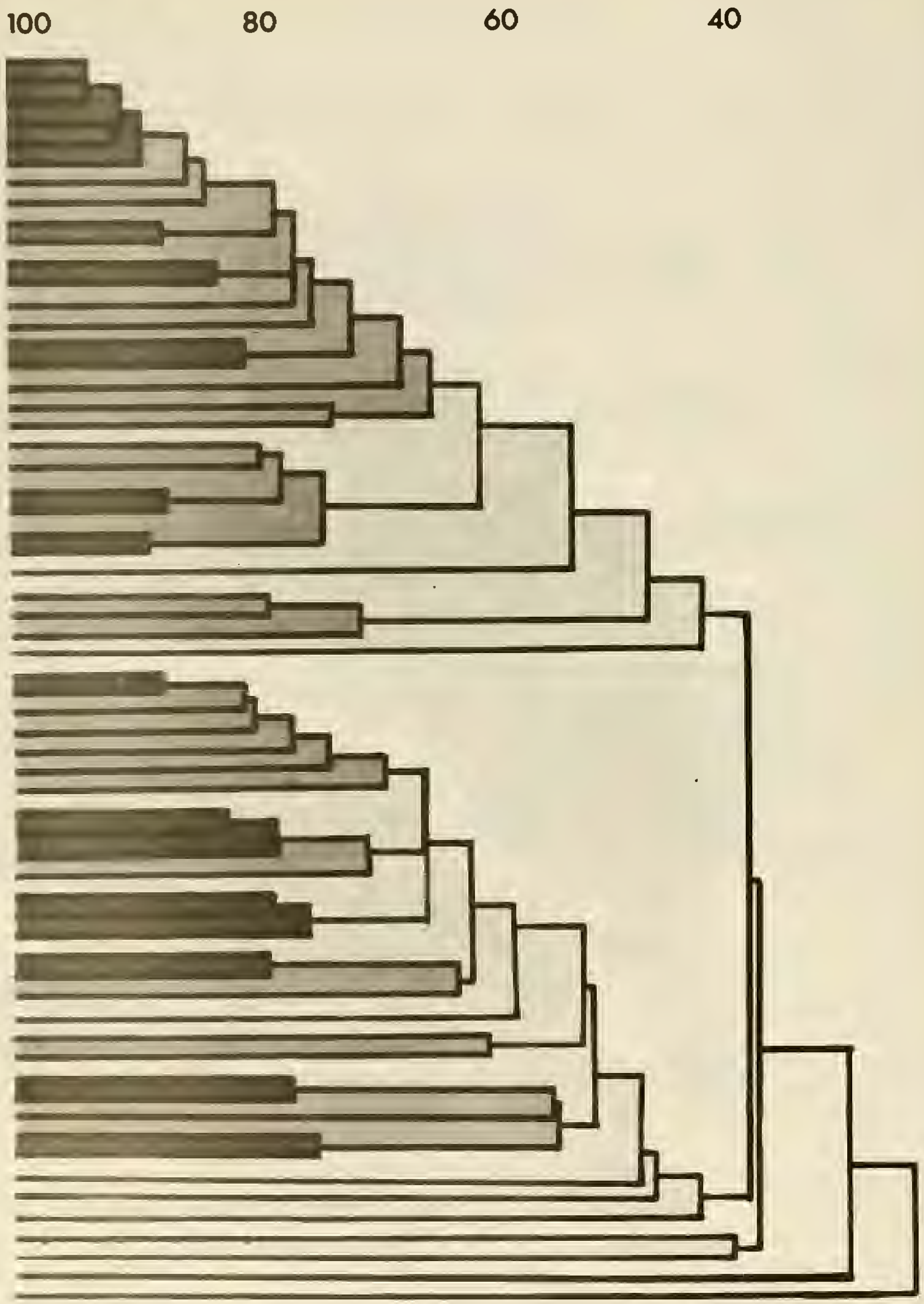

Figure 4. Biotope classification based on cluster analysis of samples; each horizontal line represents a sample; there are two major biotopes, each composed on several minor biotopes.

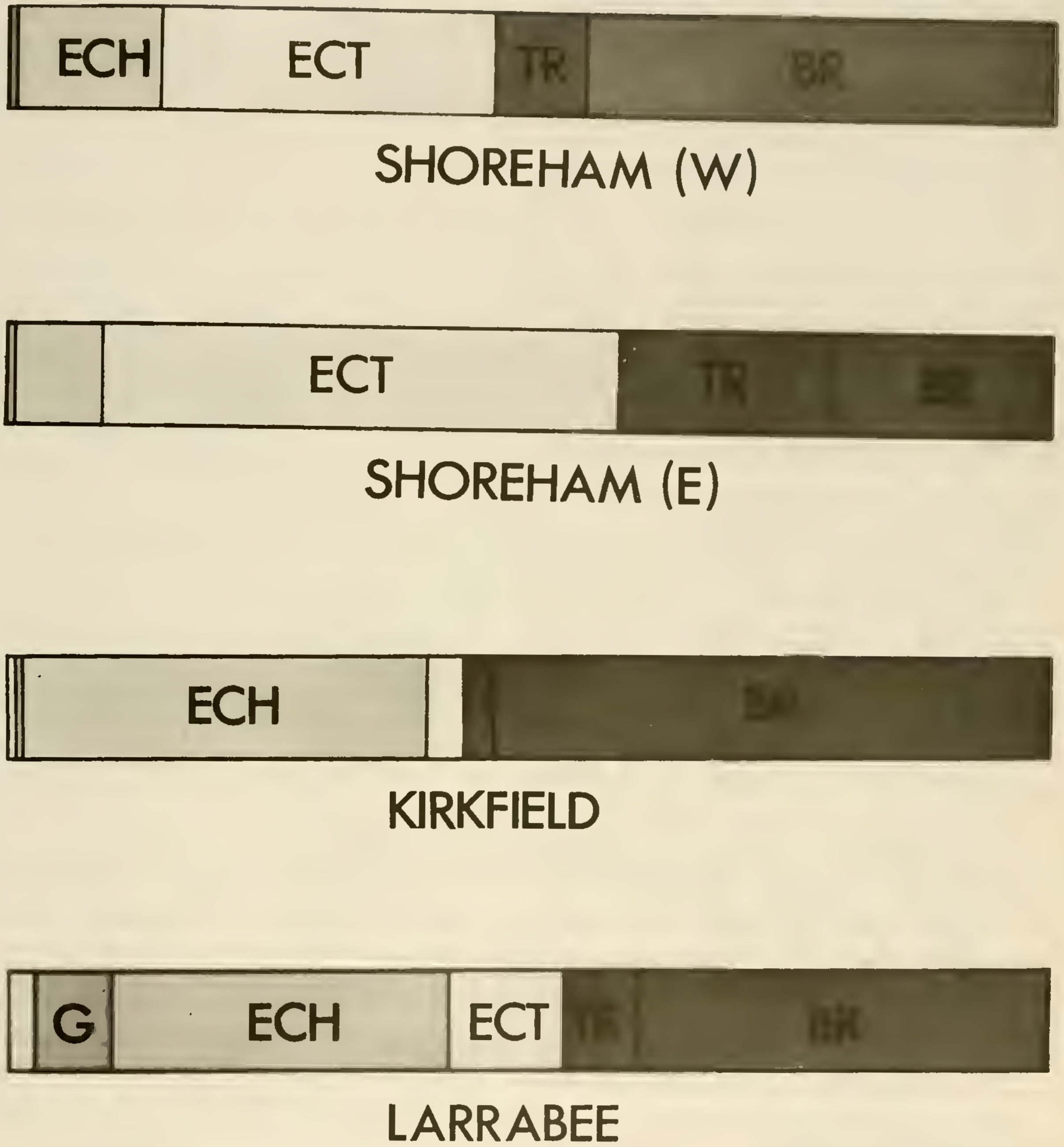

Figure 5. A comparison of the average faunal compositions east and west of the Adirondack Arch; the "Shoreham" is now considered the Sugar River Limestone west of the arch and the upper Glens Falls Limestone east of the arch; the "Kirkfield" is the Kings Falls Limestone and the Larrabee is the lower member of the Glens Falls Limestone. G=gastropod, ECH=echinoderm, ECT=ectoproct, TR=trilobite, and BR=brachiopod.

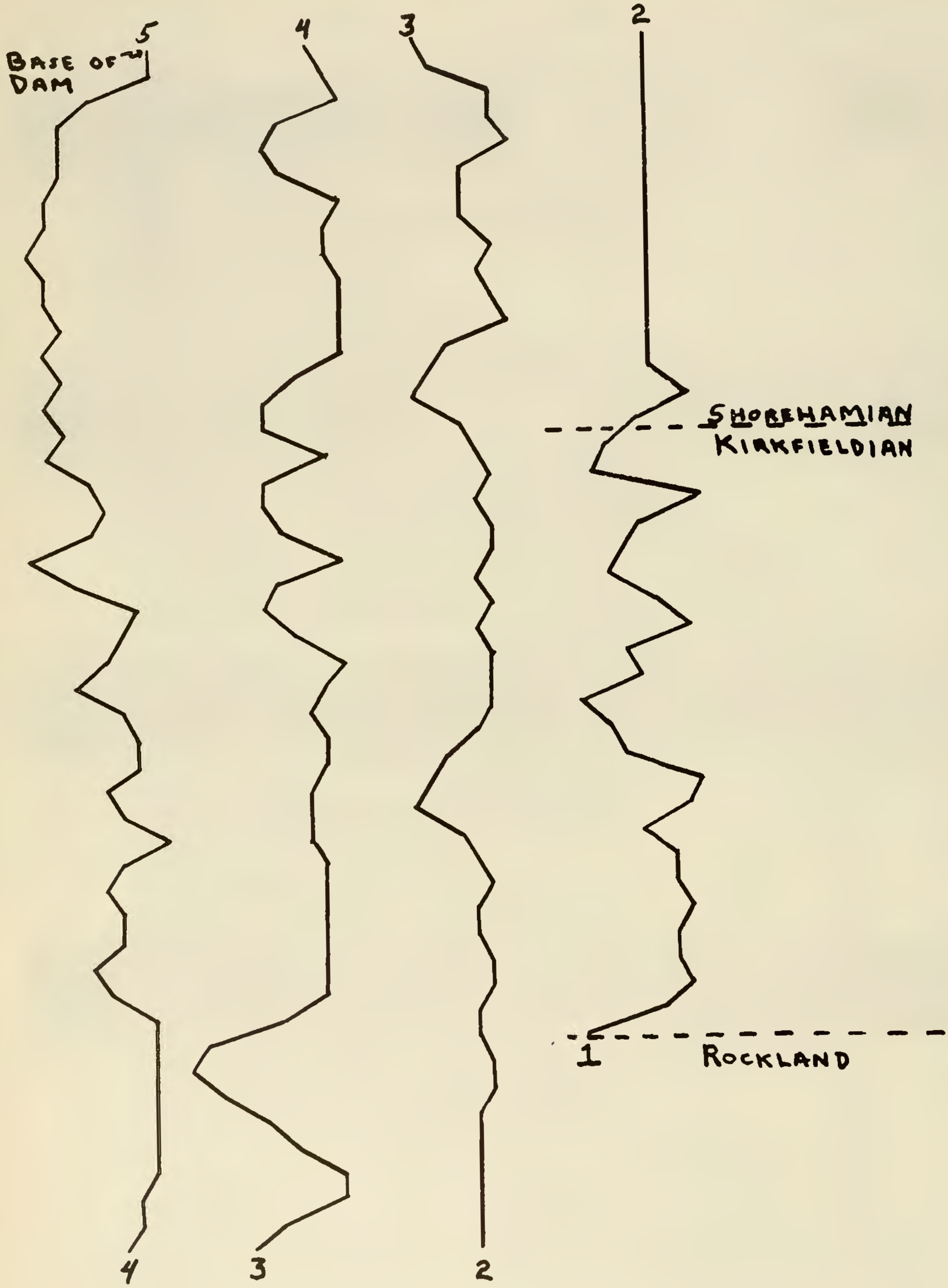

Figure 6. Time-trend analysis of Kirkfieldian and Shorehamian biofacies at Inghams Mills, New York. The line represents a smoothing of the data using a 5-term smoothing equation; it is continuous but has been plotted in 4 segments for convenience of illustration. See text for discussion.

STRATIGRAPHY OF UPPER BOLARIAN AND LOWER TRENTONIAN LIMESTONES: HERKIMER COUNTY

by

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INTRODUCTION

For almost 150 years the Ordovician rocks of the Mohawk Valley have been under study. Some of the great geologists and paleontologists of the 19th and 20th centuries have studied the limestones, shales, and fossils of the Black River Group and Trenton Group in the Black River, West Canada Creek, and Mohawk River valleys (see Kay, 1937, for historical review). As a result, these rocks have become well-known as part of the medial (middle time of a three-fold subdivision) Ordovician standard section of North America.

However, the geology of the Black River Group and Trenton Group still poses several relatively complex problems of interpretation for the application of stratigraphic, paleontologic, and paleoecologic principles. Specifically, there is still "...confusion and disagreement regarding the correlation of the upper Black River and lowest Trenton in New York and Ontario" (Kay, 1942, p. 599). Exposures are occasionally complete, but many are not. The repetitious nature of the sedimentary layers often mask significant variations critical to correct lithostratigraphic, biostratigraphic, and paleoecologic interpretations. Present and future work depends and will depend heavily on that of earlier investigators who prepared the way by determining the basic lithostratigraphic and biostratigraphic frameworks.

The modern approaches to the paleoecologic study of carbonate rocks, such as sedimentary petrography (Walker, 1968; Textoris, 1968; Cameron, 1968a), fossil community analysis (Porter and Park, 1969), population paleontology (e. g., Ross, 1967), primary sedimentary structures and trace fossils (Cameron, 1967, 1968a, 1968b; Walker, 1968, 1969), have just recently been applied with emphasis to parts of the medial Ordovician sequence of central and northwestern New York. Many previous investigators who have studied these formations have been, by necessity, primarily concerned with lithostratigraphy, such as statistical analysis of rock types (especially Chenoweth, 1952, and Lippitt, 1959), biostratigraphy and correlation, and mapping.

Although many faunal lists have been made for the New York sections (e. g., Cameron, 1968a; Fisher, 1957; Chenoweth, 1952; Kay, 1953, 1937, 1933; Young, 1943), most major fossil groups are in need of thorough restudy, using modern paleontologic approaches and techniques. However, a few groups in New York have received careful attention and revision in recent years. These include the Brachiopoda (Cooper, 1956), Ectoprocta (Ross, 1964, 1967), conodonts (Schopf, 1966), and calcareous algae (Awramik and Cameron, 1968; Cameron and Awramik, in preparation).

The long-range interests of this author center around establishing a detailed time and lithic microstratigraphic framework for the Rocklandian, Kirkfieldian, and Shorehamian stages in central and northwestern New York and southern Ontario. This would form the basis and provide the confidence for reconstructing the environments of deposition and determining the paleogeograph. At present, special emphasis is being placed on statistical analysis of the rock types both from detailed field measurements and carbonate petrography, small scale physical and biological correlation, primary sedimentary structures trace fossils, calcareous algae, fossilization, and fossil community analysis. This information should better document the initial and subsequent wider transgression of the Trentonian sea.

The purposes of this field trip to the upper Black River Group and lower Trenton Group limestones of Herkimer County, New York, are to:

1) Demonstrate the stratigraphic succession and its lateral variations.
2) Discuss and evaluate the age relationships and time correlations of the various formations by
   a) Examining the diverse faunas and
   b) Demonstrating the lateral continuity of major lithic and biologic characteristics.
3) Examine and evaluate the criteria for determining the extent and significance of the disconformity along the Black River-Trenton boundary.
4) Examine and evaluate the criteria for determining the conditions and environments of deposition and paleogeography.

This field trip guide will summarize previous work on the Black River Group and Trenton Group in the Little Falls and Utica 15' quadrangles of Herkimer County and incorporate new data insupport of preliminary reinterpretations of the stratigraphy of part of the lower Trenton Group in this area The order of localities has been chosen as conveniently as possible for econom of travel along a southeast to northwest traverse.

GEOLOGIC SETTING

The Little Falls and Utica quadrangles are located along the southwestern margin of the Adirondack Mountains and include part of southern Herkimer County (Fig. 1). Good exposures of medial Ordovician limestones are to be found along the Mohawk River, West Canada Creek, and East Canada Creek valleys and those of their tributaries (Fig. 1). Many small abandoned limestone quarries in both quadrangles contain exposures of the Black River-Trenton boundary.

Lower Paleozoic strata dip gently to the southwest from the Precambrian on the northeast; subsurface contours drawn on the base of the Black River-Trenton indicate a $1\frac{1}{2}$ degrees dip southwestward (Flagler, 1966, pl. 5). Precambrian rocks are exposed over much of the northeastern part of the Little Falls Quadrangle (Cushing, 1905a) and a Precambrian inlier occurs at Middleville (Fig. 1; Cushing, 1905a; Young, 1943; Kay, 1953). A few northeast-southwest trending normal faults cut Paleozoic and Precambrian rocks, e. g. near Little Falls and Dolgeville (Cushing, 1905a; Kay, 1937).

The late Cambrian Little Falls Dolomite underlies Ordovician rocks and overlies the Precambrian basement complex of igneous and metamorphic rocks in southern Little Falls Quadrangle. To the northeast, however, the medial Ordovician Gull River Limestone overlaps the Little Falls Dolomite to lie directly on the Precambrian (Cushing, 1905a; Young, 1943), as it does farther

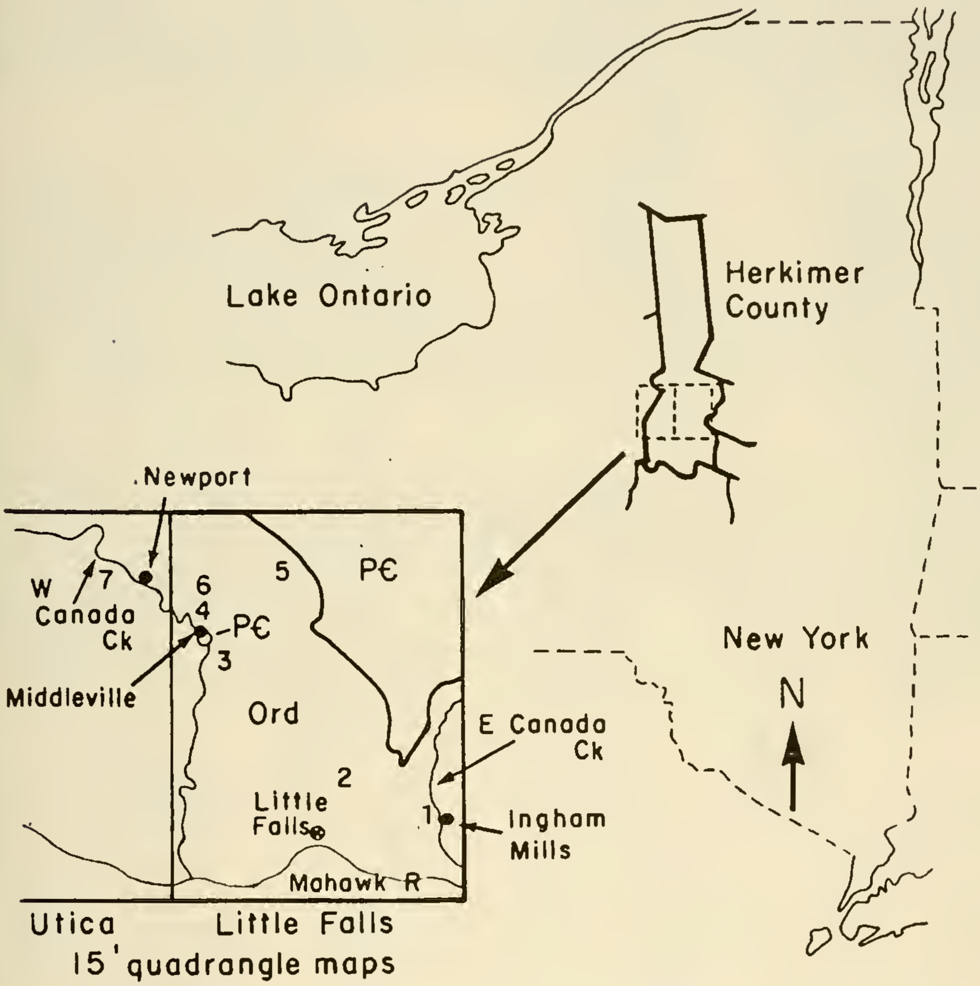

Fig. 1. Partial map of New York State with index maps showing location of quadrangles and field trip stops (numbers).

north in Oneida County (Young, 1943, fig. 3). The Ordovician rocks of the Little Falls and Utica quadrangles occur in the southern third of Herkimer County (Fig. 1). Silurian and Devonian rocks comprise the bedrock geology of southernmost Herkimer County (Rickard and Zenger, 1964).

The Ordovician rocks of the Little Falls and northeastern corner of the Utica quadrangles range from the early Ordovician (Canadian) limestones and dolostones of the northward disappearing Tribes Hill and Chuctanuda formations (Fisher, 1954) through the medial Ordovician Gull River Limestone of the Black River Group to the medial Ordovician limestones and shales of the Trenton Group. Pleistocene and Recent sediments cover much of the northern area (Cushing, 1905a; Kay, 1953).

STRATIGRAPHIC CLASSIFICATION AND PROBLEMS

Some controversy exists today as to whether the rock units of the Black River Group and Trenton Group in New York and Ontario are time-stratigraphic, i. e., the same age throughout their geographic distribution (Kay, 1937, 1942; Young, 1943; Chenoweth, 1952; Cameron, 1968a, 1969), differ in age (Sinclair, 1954), or vary to the extent of being time-transgressive (Winder, 1960; Fisher, 1962; Barnes, 1965, 1967).

Early workers applied a single term for the rock units and the time-rock units (stages) because they were principally interested "...in distinguishing rocks of an age rather than of one kind" (Kay, 1968b, p. 1373). These time-rock units were called stages by Clarke and Schuchert (1899), Cushing (1905b), and Kay (1937, 1947), but others (Grabau, 1913; Willis, 1901) generally called them formations, relying upon context for distinction between the two concepts (Kay, 1968b). "Formations formed divisions of the 'standard time-scale' (Williams, 1901, p. 573)" (Kay, 1968b, p. 1373).

The three terms Pamelia, "Lowville", and "Chaumont" have been used interchangeably as rock and time-rock units for the Black River Group. Clarke and Schuchert (1899) proposed and used the "Lowville" as a stage, while Kay (1929, p. 664) proposed the "Chaumont", as a time-stratigraphic term, for rocks "...younger than Lowville and older than the Rockland." The terms Chaumontian and Lowvillian, with the suffix "-ian" added to clarify the author's intent, are used as stage names (Fig. 2). The outcrop belt of these units in northwestern New York may actually parallel the original shoreline, suggesting that they are most probably both lithic and time-stratigraphic in nature. Walker's (1969) work in the Black River Valley west of the Adirondacks suggests this is for at least the upper half of the Black River Group.

Raymond (1914, 1921) named the "Rockland" and succeeding Trenton formations and "...applied the names to units recognized by a succession of faunal zones; these were chronostratigraphic" (Kay, 1968a, p. 167). Thus, biostratigraphic units came to be used as lithostratigraphic units which seem to paralle time lines independently drawn from studies of the faunas and the lithologies, including metabentonites (Kay, 1937, 1942, 1943, 1953; Young, 1943; Chenoweth, 1952; Lippitt, 1959; Cameron, 1968a, 1969). These lithic units are not completely uniform throughout, but change systematically along the outcrop belt and contrast with overlying and underlying units in a constant way (Chenoweth, 1952; Cameron, 1968a).

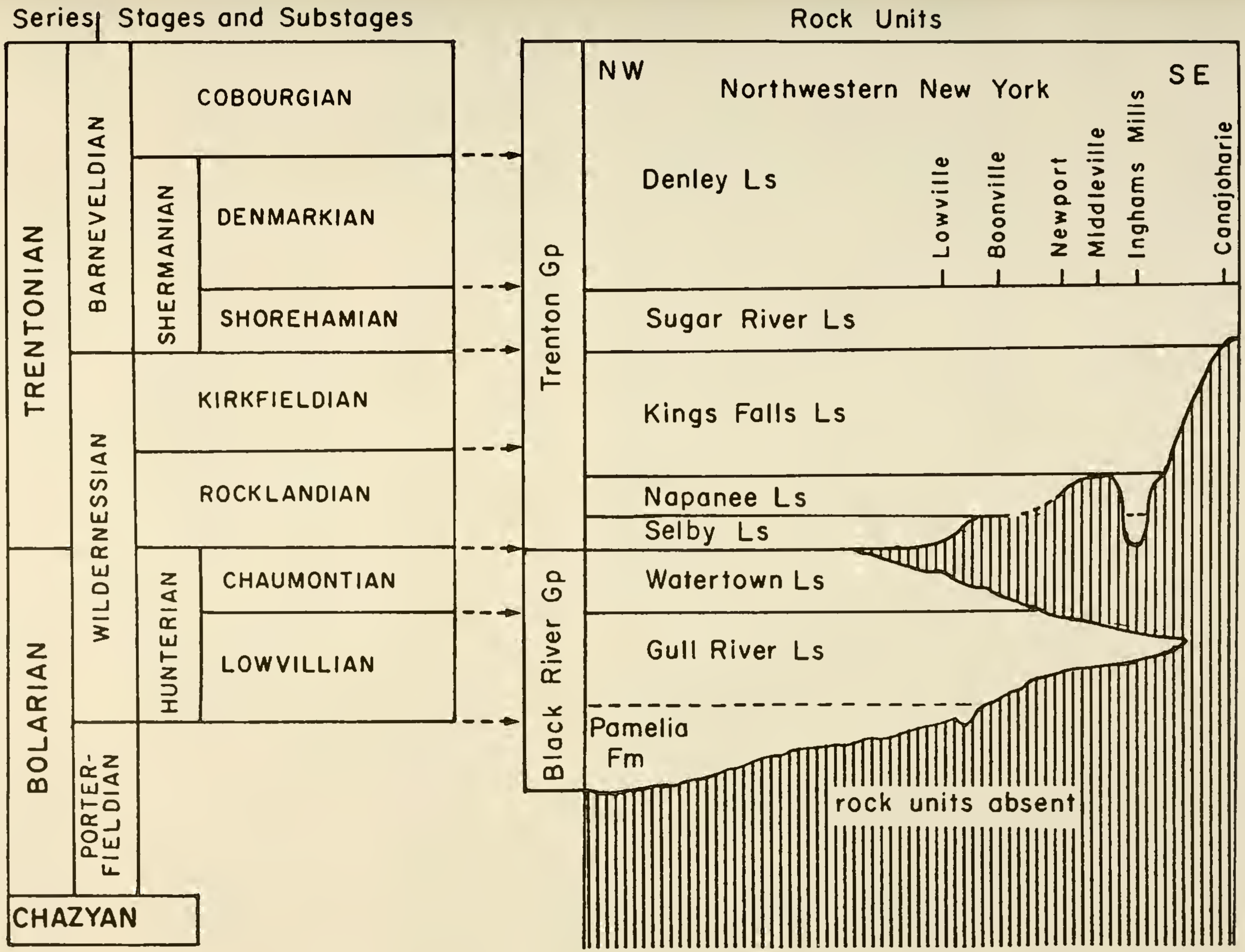

Fig. 2. Medial Ordovician stratigraphic classification and nomenclature for northwestern New York.

Fisher (1962) stated that "...misunderstanding persists, owing to mixed usage of lithostratigraphic, biostratigraphic, and chronostratigraphic units - all termed 'formations'. Furthermore, some geographic names are used in a dual or even triad sense (viz, Rockland Limestone, Rocklandian Stage, Rocklandian-*Dalmanella* zone), and one is never quite certain of the writer's intention." According to standard usage, Rockland Limestone ought to be a lithic unit, Rocklandian Stage a time-stratigraphic unit, and *Dalmanella* zone a biostratigraphic unit. To remedy the dual usage of nomenclature, some paleontologists and stratigraphers have been adding the suffix "-ian" or "-an" to the formation names to clearly indicate whether a stage, i. e., a time-rock unit is being discussed. Rocklandian and Kirkfieldian (Fig. 2) were used as stages by Kay in 1935, although not with the "-ian" ending until 1948 when he used the name Trentonian, as did Grabau (1909).

Many paleontologists, working on different fossil groups independently, believe that some other area than New York should be used as the medial Ordovician standard (Cooper, 1956; Fisher, 1962). However, not all paleontologists and stratigraphers agree with this viewpoint (e. g., De Mott, 1963). Cooper (1956) using mainly brachiopods considered the Black River no longer useful because it was so defined (see Kay, 1937) that it did "...not describe the stratigraphy..." and "...the interval represents only a part of what we regard as a natural grouping of faunas" (p. 7). He believed the Black River faunas were closely related to the faunas of the lower Trentonian limestones and rejected Kay's (1947) Bolarian Series. He proposed the "Wilderness" as a stage term for the upper Bolarian and lower Trentonian interval and restricted the "Trenton" to the medial and late Trentonian of Kay's (1937) usage. Fisher (196 substituted "Barneveld" for this restricted "Trenton" Stage and kept the terms Trenton and Black River for rock-stratigraphic units (groups). Cooper overlooked the Kirkfieldian Stage which Fisher (1962) added to the top of the "Wilderness".

Kay (1929, *et sub. pub.*), Young (1943), Chenoweth (1952), Lippitt (1959), Cameron (1968a, 1969), and Walker (1969) have mapped and described comprehensively much of the stratigraphy of Bolarian and Trentonian limestones in northwestern New York. The rank of some units, such as the Watertown, Selby, and Napanee, have been changed for geologic and nomenclatural reasons (Cameron, 1967, 1968a; Kay, 1968a, 1968b). Kay (1968b) extended from southern Ontario the use of Okulitch's (1939) term Gull River Limestone for the similar limeston of the Black River Group in northwestern and central New York below the Watertown Limestone and above the Pamelia Formation (Fig. 2), so that Lowvillian could be used exclusively as a time-rock term. Watertown Limestone (Cushing, *et al.*, 1910) is used for the Bolarian limestones above the Gull River Limestone (Kay, 1968a, 1968b). Kay (1968b) also named and defined younger formations (Kings Falls Limestone, Sugar River Limestone, Denley Limestone) of the Trenton Group in order to have completely separate and unambiguous sets of time and rock nomenclature (Fig. 2).

The stratigraphic classification used herein (Fig. 2) follows that of Kay (1968b) with modifications for the lower Trenton Group from Cameron (1967, 1968a, 1969). A thorough historical review of the early classifications of these limestones can be found in Kay (1937, p. 237-249); for a review of later work, see Cameron (1968a).

UPPER BOLARIAN AND LOWER TRENTONIAN STRATIGRAPHY OF HERKIMER COUNTY

Introduction:

The medial Ordovician Gull River Limestone, "Napanee Limestone", Kings Falls Limestone, and Sugar River Limestone of the Little Falls and Utica 15' quadrangles are described in their order of deposition (Fig. 2). Their age and stratigraphic relationships are emphasized.

Bolarian Series:

The Gull River Limestone was deposited under very shallow marine conditions during the Lowvillian Stage; apparently neither Chaumontian (uppermost Bolarian) nor pre-Lowvillian rocks are present in this region. The Bolarian age is indicated by conodonts (Hasan, 1969). The Gull River lies successively on early Ordovician (Canadian) limestones and dolostones along the Mohawk River, on late Cambrian Little Falls Dolomite farther north, and on the Precambrian along its northerly outcrop belt (Cushing, 1905a; Young, 1943). The thickness of this formation varies in a southeastwardly direction from 40 feet at Newport to 27 to 30 feet at Inghams Mills (Figs. 2 & 3).

The lithology of the Gull River is varied and complex, with two unnamed members recognized in the Utica and western Little Falls quadrangles. The lower subdivision is composed of tan weathering, gray, medium-textured, heavy-ledged, tough, quartz arenite-rich, argillaceous, vertically burrowed limestone with thin shale layers. A 3-inch thick metabentonite (altered volcanic ash) occurs near the top at City Brook (Stop #4) which may correlate with the base of the upper member at Newport (Kay, 1953). The lower member varies from 8 to 14 feet thick, being thickest to the north (Fig. 3). The thicker upper subdivision (19 to 30 feet) is more widespread and characterized by light gray weathering, dove gray calcilutite (sublithographic), called "birdseye" limestone by the early geologists in New York State. Granule and flat pebble calcirudites and impure argillaceous calcisiltites are sometimes frequent. Horizontal laminae, fenestral fabric, stylolites, mudcracks (Fig. 5), and burrows are common sedimentary structures. Some of the horizontal laminae originated by current action, but most probably were produced by algal accretion (algal? mats).

Fossils are generally uncommon in the Gull River Limestone (see Young, 1943, for a comprehensive faunal list). The abundant vertical burrow Phytopsis (Fig. 7) and small ostracods occur throughout the Gull River. The large ostracod Eoleperditia fabulites, snails, trilobite fragments, cryptostome bryozoa, and the tabulate coral Tetradium cellulosum are relatively common in the upper third of this formation. T. cellulosum (Fig. 8) and the vertical burrow Phytopsis are characteristic of the Gull River and are often abundant.

Trentonian Series:

Rocklandian Stage. Two limestone units in Herkimer County are thought to be lowest Trentonian in age: (1) the "Napanee Limestone" in the Inghams Mills area (Stop #1) and (2) the black chert-bearing "calcisiltite lithofacies" well-exposed in the vicinity of Newport (Stops #5-7). These units occur inbetween the Gull River and Kings Falls limestones which lie directly in contact with each between Inghams Mills and Middleville (Figs. 1-3).

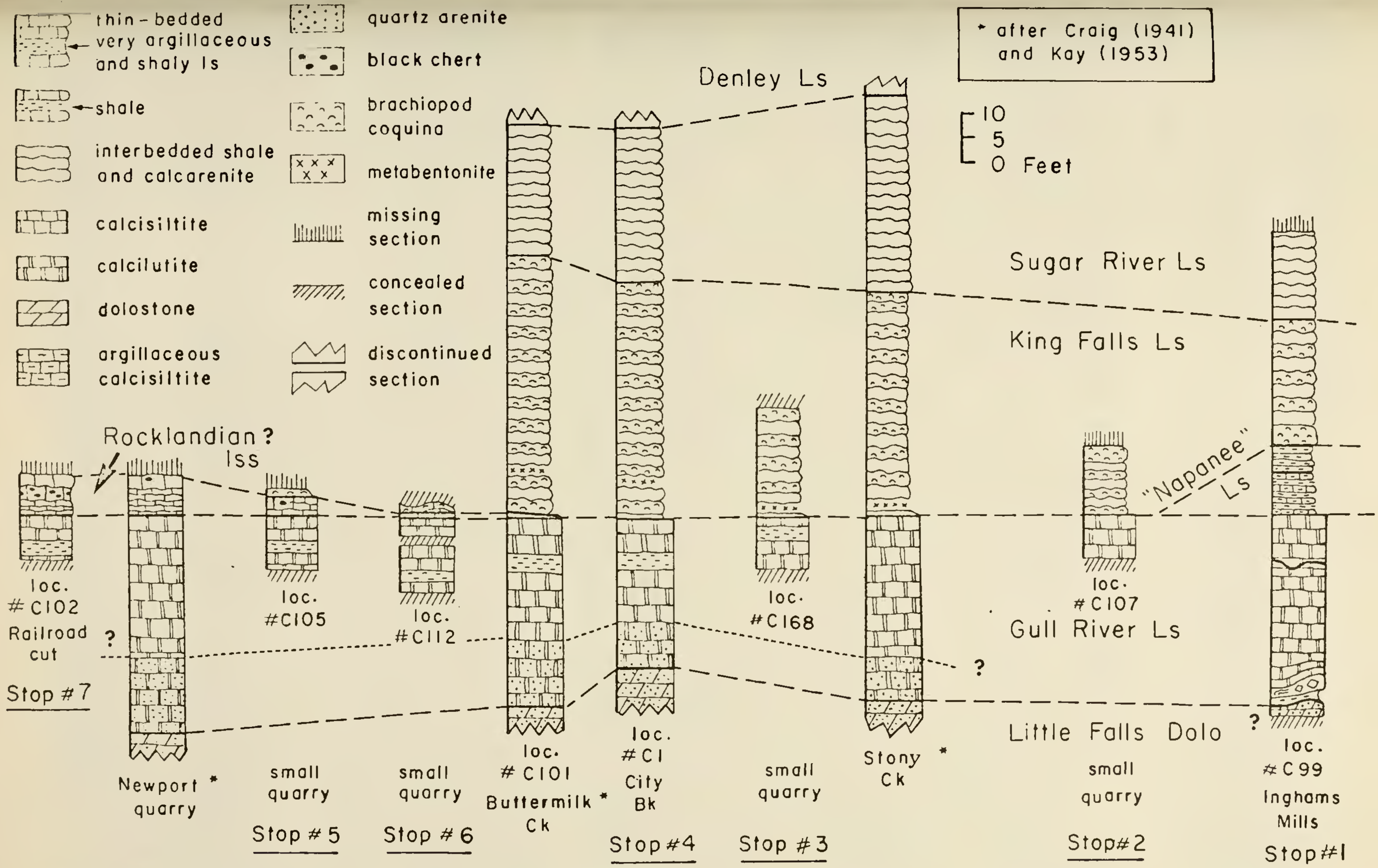

Fig. 3. Measured sections of upper Bolarian and lower Trentonian limestones between Newport and Inghams Mills, New York.

The black chert-bearing "calcisiltite lithofacies" is composed of medium gray weathering, dark gray to black, generally medium-bedded and heavy-ledged, irregularly burrowed, argillaceous, brittly fracturing, sparsely fossiliferous micrites. A few fossiliferous biomicrites and biosparites are occasionally present. Black chert nodules occur frequently in the thicker northerly sections near Newport where thin shale layers begin to appear in the middle of the facies. Burrows occur as interconnected non-vertical and vertical burrows resembling Camarocladia of the upper Bolarian Watertown Limestone ans as burrow reworking which has apparently destroyed any original current laminations. "Corrasion" surfaces mark its contacts with the Gull River and Kings Falls limestones (Kay, 1953, fig. 28). Maximum thickness of this facies is about 7 to 8 feet in the vicinity of Newport (Stop #7), but it decreases to zero southward 2 miles north of Middleville (Stop #4). Due to concealment by Pleistocene sediments, no exposures of this interval can be found north of Poland for 23 miles until Boonville where the Rocklandian Napanee Limestone is present at this stratigraphic level. Although few time-diagnostic macrofossils (Rafinesquina reported by Craig, 1941, and Kay, 1953) have been identified from the "calcisiltite lithofacies". conodonts (Hasan, 1969) indicate a Trentonian age.

Thirteen feet of interbedded calcisiltite and calcareous shale, divisable into two unnamed members, comprise the lowest Trentonian limestones in the Inghams Mills area (Stop #1; Figs. 7-8; Kay, 1937, pl. 3, fig. 1). The lower member ($7\frac{1}{2}$ feet) is composed of chocolate brown weathering interbedded shales and argillaceous, burrowed, black calcisiltites, while the upper member ($5\frac{1}{2}$) is composed of medium gray weathering, dark gray to black, less argillaceous, burrowed calcisiltites (micrites and biomicrites) and thinner interbedded calcareous shales. Horizontal and cross-laminations are absent, probably due to complete burrow reworking. Skeletal calcarenites (biosparites) are sparse but increase in frequency towards the top of the upper member. The surfaces of these limestones exhibit extremely well-developed loading casts, suggesting that the weight of overlying sediments had deformed these limestones because they were still incompletely consolidated after burial.

Because these Rocklandian limestones (especially the upper member) resemble the upper Rocklandian Napanee Limestone of northwestern New York in that they are composed of interbedded calcisiltites and shales, they are tentatively referred to as "Napanee Limestone". The Napanee also outcrops in the southern Adirondacks, present in the infaulted outlier at Wells where about 10 feet are exposed (Fisher, 1957). Elsewhere in the Mohawk Valley of central New York Rocklandian limestones are absent.

Fossils are uncommon in the lower "Napanee Limestone" which is characterized by the trilobites Isotelus and Bumastus and the ectoproct Batostoma. Fossils are more abundant and diversified at the top of the lower member and especially in the upper member (Cameron, 1968a). The Rocklandian age is indicated by the lower Rocklandian Doleroides ottawanus Zone in the lower member, the upper Rocklandian Triplesia cuspidata Zone in the upper member, and by conodonts (Schopf, 1966; Hasan, 1969).

Kirkfieldian Stage. The widespread Kings Falls Limestone is characterized by coarse-grained, relatively thick-bedded, coquinal calcarenites (Fig. 3), and is primarily composed of dark gray weathering, dark gray to black, argillaceous, fine- to coarse-grained, coquinal and non-coquinal calcarenites with interbedded thin calcareous shales. Calcisiltites are uncommon and calcilutites are rare. Pararipples (Fig. 12) and cross-laminated beds are common (see Chenoweth, 1952). In this region the lower half is apparently thinner bedded than the upper half (Craig, 1941; Kay, 1953).

The thickness of the Kings Falls varies from 40 to 45 feet in the vicinity of Newport and Middleville and progressively thins southeastward, being 23 feet thick at Inghams Mills. Farther east in the Mohawk Valley it disappears before Canajoharie. However, Fisher (1957) questionably recognized a few feet of "...subcoquinites and calcarenites...", typifying this unit, in the Wells outlier of the southern Adirondacks (p. 17). To the northwest, west of the Adirondacks, it increases to 69 feet near Boonville and 100 feet farther north along the Black River Valley (Chenoweth, 1952).

In the vicinity of Middleville, a thin metabentonite (altered volcani ash) occurs at 9 feet above the base along Buttermilk Creek, 7 feet along City Brook (Stop #4), one foot (possibly another at 2 feet) in the small quarry (Sto #3) southeast of Middleville, and at 2 feet along Stony Creek (Fig. 3). If this persistent clay represents a synchronous time surface at each of these localities, then the base of the Kings Falls is onlapping the Gull River and becoming progressively younger eastward.

The fauna of the Kings Falls Limestone has not been studied in detail but "...the brachiopods Dalmanella rogata (Sardeson) and Sowerbyella punctistria (Mather) and Rafinesquina trentonensis Salmon and the bryozoan Prasopora sp. ar common. At Inghams Mills, in the Little Falls quadrangle, the writer recognize the following genera on a single bedding surface; Solenopora, Streptelasma, Rhinidictya, Dalmanella, Dinorthis, Hesperorthis, Opikina, Parastrophina, Sowe byella, Hormotoma, Liospira, Phragmolites, Subulites, "Cyrtoceras", Geisonocera (?), Bathyurus, Calliops, Ceraurus, Encrinurus, Hemiarges and Leperditia" (Kay, 1953, p. 43).

Shorehamian Stage. The Sugar River Limestone is composed of relative thin-bedded (generally contrasting with the heavier-ledged upper beds of the Kings Falls), dark gray to black, very argillaceous, generally coarse-grained, non-coquinal calcarenites interbedded with calcareous shales. Along West Canada Creek, the upper 5 to 9 feet consist of calcilutitic limestones of the Rathbun member (Chenoweth, 1952; Kay, 1953) which will not be seen on this field trip. The pre-Rathbun Sugar River limestones are thicker in northwestern New York (Chenoweth, 1952) and decrease in thickness eastward from 40 to 17 feet in western Herkimer County. They persist eastward in the Mohawk Valley and almost span the Adirondack Arch (Kay, 1937; Fisher, 1954), being 17 feet thick at Canajoharie.

Lithically, the Sugar River limestones are easily distinguished from the subjacent Kings Falls limestones in Herkimer County by the sparseness of brachiopod-rich coquinal calcarenites, which characterize the Kings Falls, and the greater abundance of coarse-grained, non-coquinal calcarenites. Calcareous shales are generally more abundant in the Sugar River, especially in northwestern New York (Chenoweth, 1952), but at Inghams Mills shales appear to be less abundant than in the subjacent Kings Falls.

The fauna of the Sugar River is described and listed in Kay (1937, 1953) and Chenoweth (1952). The Cryptolithus tesselatus Zone occurs throughout this unit. The trinucleid trilobite C. tesselatus Green "...is usually accompanied by Trematis terminalis (Emmons) and Prasopora..." simulatrix (Kay, 1937, p. 267). The brachiopods Dalmanella, Sowerbyella, and Rafinesquina and the trilobites Isotelus and Flexicalymene are common. However, brachiopods are less common than in the Kings Falls limestones below, while cryptostome bryozoa are more common.

DESCRIPTIONS OF INDIVIDUAL STOPS

Stop #1. Inghams Mills (locality #C99):

Because four formations of medial Ordovician age are excellently exposed along with the top of the Little Falls Dolomite(?) below the dam on East Canada Creek, much time and attention will be given to this outstanding outcrop (Figs. 3-12). Lithologies, unusual sedimentary structures, fossils, formation boundaries, and two disconformities will be carefully examined.

Little Falls Dolomite(?). Only two feet of the late Cambrian Little Falls Dolomite are exposed at the base of the exposure. This unit is represented by relatively thick-bedded, light to medium brown weathering, quartz arenitic, pyrite-bearing dolostone with thin interbedded shale layers.

Gull River Limestone. About $29\frac{1}{2}$ to 30 feet of Gull River Limestone are excellently exposed with a thick dove gray shaly limestone at the base. The lowest two to three feet exhibit a slump breccia (Fig. 4), containing limestone blocks up to $2\frac{1}{2}$ feet in diameter, that probably formed as a result of instability over the irregular depositional surface of the Little Falls Dolomite.

The next $16\frac{1}{2}$ feet contain horizontally laminated (algal?), dove gray calcilutites with abundant vertical burrows (Phytopsis), a few ostracods, and frequent stylolites. Frequent mudcracks confirm an intertidal or lagoonal origin.

Thin shales are common in the lower 10 feet. The folded limestone layers reported by Cushing (1905a, pl. 6) from the lower Gull River Limestone at this exposure apparently formed as a result of settling over compacting thick shale lenses (Fig. 6).

Between $16\frac{1}{2}$ and $21\frac{1}{2}$ feet an apparently subtidal, irregularly burrowed, essentially non-laminated, massively bedded, dark gray to black calcisiltite zone contains Foerstephyllum halli, Lambeophyllum profundum, Hormotoma, Loxoplocus, Isotelus, cryptostome bryozoa, straight nautiloids, and pelmatozoan debris.

Immediately above these deeper water sediments, the intertidal or lagoonal facies begins to reappear. This is a vertically burrowed, horizontally laminated (algal?), limestone intraclast-bearing, fossiliferous calcilutite and calcisiltite zone. Fossils from this interval include Tetradium cellulosum, Eoleperditia fabulites, Lambeophyllum profundum, Isotelus, cryptostome bryozoa, and pelmatozoan fragments. Near the base of this zone a sediment-filled tidal meander(?) or channel (Fig. 7) up to 7 feet wide and 2

Stop #1. Inghams Mills

Fig. 5. Mudcracks in Gull River Limestone.

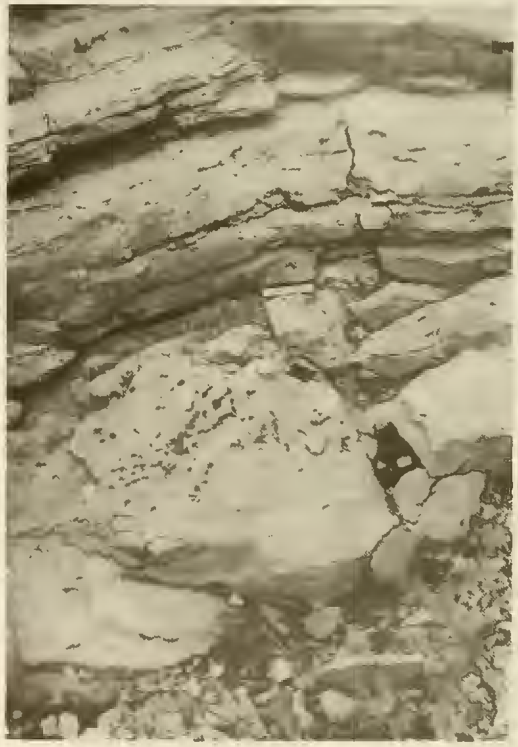

Fig. 4. Slump breccia at base of Gull River Limestone. (Scale is shown by 6-inch ruler in all photographs.)

COMPACTION FOLDING

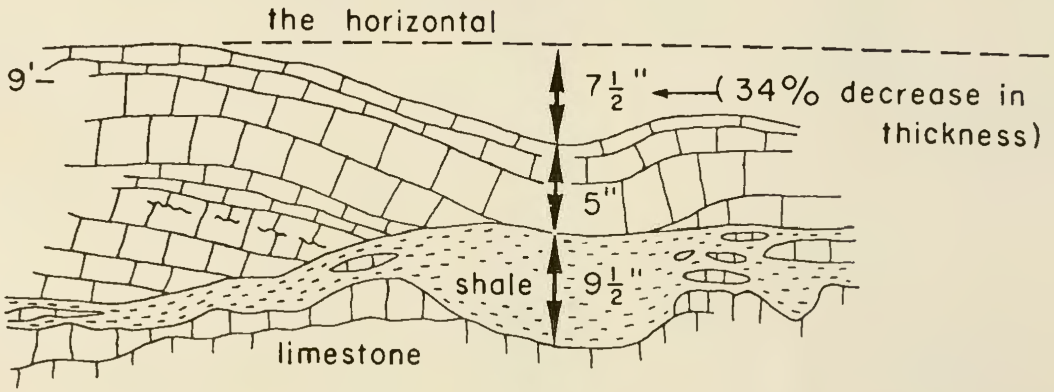

Fig. 6. Fold developed in Gull River limestones at Inghams Mills (Stop #1) that probably formed as a result of settling over a compacting thick lens of shale. Such folding has produced broad anticlines (depending upon the spacing of underlying shale lenses) and narrow synclines. The decrease in thickness of the section over the syncline provides a quick, easy measure of the minimum percentage decrease in thickness of the shale lens due to compaction. The shale lens shown in the diagramatic field sketch above was probably about 17 or more inches thick originally.

feet deep is excellently exposed in two faces of the outcrop. Note the structure and composition of the sediments filling it.

At about 27 feet, a 9-inch thick calcilutite bed contains scores of whole *Tetradium cellulosum* colonies in life position, possibly representing an inner shelf reef, as suggested by Fisher (1965) for occurrences in the eastern Mohawk Valley. They cover 50% to 90% of the bed which contains a thin veneer of limestone pebble conglomerate. One can readily see how the fine-grained sediment was trapped in and around these delicately branching tabulate corals.

The top of the Gull River Limestone is riddled with burrows (dominan tly vertical) partially filled with the black lustrous carbonaceous mineral anthraxolite. Several inches of irregular relief over the top of this bed (Fi marks the disconformity between the Black River Group and the Trenton Group.

"Napanee Limestone". The lowest 13 feet of the Trenton limestones can be divided into $7\frac{1}{2}$ feet of chocolate brown weathering interbedded calcareo shales and argillaceous calcisiltites at the base and $5\frac{1}{2}$ feet of medium gray weathering interbedded thinner shales and less argillaceous calcisiltites above. The contact between these two subdivisions is slightly gradational. The surfaces of the limestone layers exhibit extremely well-developed loading casts. In addition, the lower subdivision contains an unusually well-develope and fully exposed intraformational (subsolifluction) fold (Fig. 10) similar to those described by Chenoweth (1952) from the Sugar River Limestone in nort western New York. Fossils are common.

Kings Falls Limestone. Contrary to previous interpretations (Kay, 1937, 1953; Fisher, 1954; Cameron, 1968a), 23 feet of Kings Falls Limestone are exposed above the "Napanee Limestone". Polymictic (dominantly limestone) conglomerates (Fig. 11) and coquinal calcarenites mark its base. The boundary with the Sugar River Limestone above is marked by a rather sharp decrease in brachiopod-rich coquinal calcarenites which give way to an increase in encrinitic bryozoan-rich, coarse-grained calcarenites. Cross-laminated beds are common, as are large ripples (pararipples) whose average ripple spacing is 25 to 26 inches and average ripple height is 1.9 to 2.3 inches. Fossils are abundant.

Sugar River Limestone. About 15 feet of dominantly non-coquinal calcarenite with interbedded thin calcareous shales comprise the Sugar River Limestone immediately below the dam. Several beds of limestone pebble conglomerate and a mudcracked layer occur near the base at the center of the exposure. Cross-laminated beds and ripples are frequent and mudcracked beds also occur near the top.

The Shorehamian *Cryptolithus tesselatus* Zone is present at this interval where *C. tesselatus* has been found in the upper 8 feet and abundant *Prasopora* sp. dominate the upper 13 feet. The top of this exposure may actual be close to the top of the pre-Rathbun Sugar River Limestone because unusuall large *Prasopora* occur here in the upper few feet as they do in the upper few feet of this formation along City Brook (Stop #4). Fossils seem less abundant than in the Kings Falls Limestone below because of greater shell comminution.

The fossils, lithologies, and sedimentary structures at this localit suggest decreasing water depths in this area during the lower Trentonian. Certainly, the greater shell comminution and mudcracks of the Sugar River lime

Stop #1. Inghams Mills

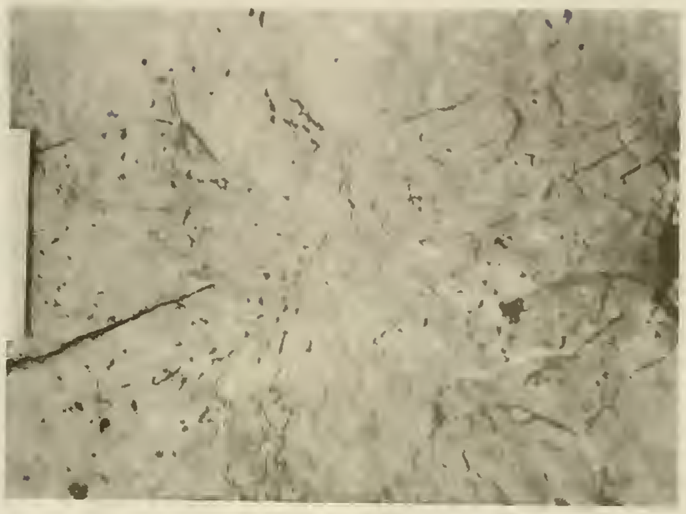

Fig. 8. Top view of whole Tetradium cellulosum colony from top of Gull River Limestone.

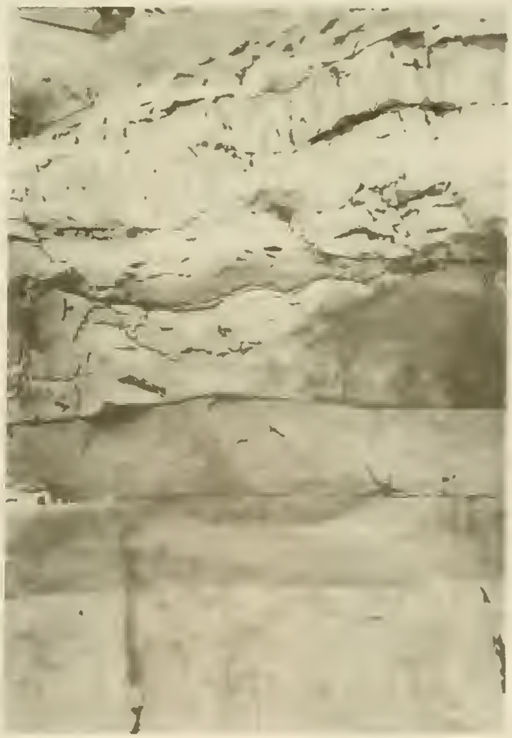

Fig. 7. Upper Gull River Limestone showing vertically burrowed (Phytopsis) beds, darker subtidal facies, and channel deposits.

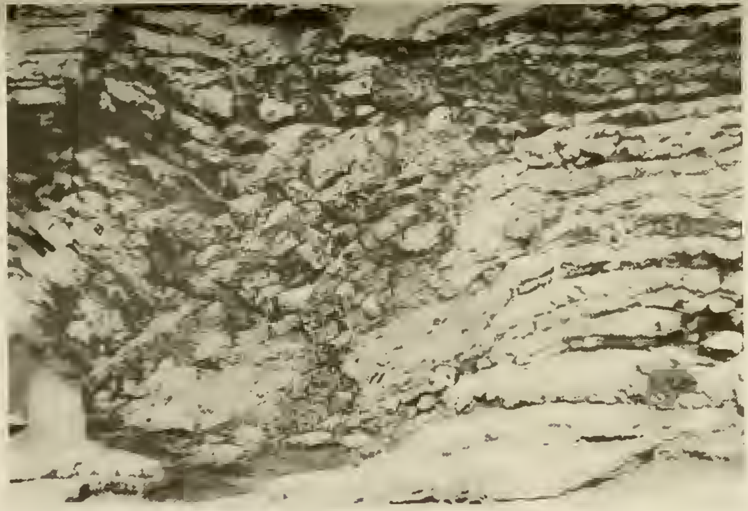

Fig. 10. Subsolifluction fold in lower "Napanee Limestone".

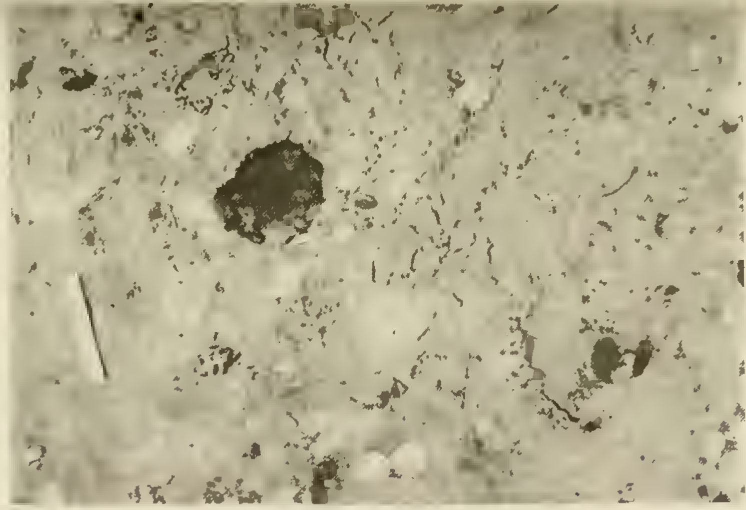

Fig. 11. Limestone conglomerate at base of Kings Falls Limestone.

Fig. 9. Disconformity between lower "Napanee" and Gull River limestones.

Fig. 12. Pararipples in lower Kings Falls Limestone.

stone indicate shallower water during the Shorehamian than during the Kirkfieldian Stage. The older Rocklandian at this locality may have been deposited either under somewhat restricted conditions or at greater depths.

<u>Stop #2</u>. Small quarry (locality #C107):

The Black River-Trenton boundary is well-exposed along the contact between 8 feet of upper Gull Limestone and 13 feet of lower Kings Falls Limestone (Fig. 3). The "Napanee Limestone" has disappeared and intraclasts of the Gull River can be seen in the lowest few inches of the Kings Falls. These intraclasts may indicate that the Gull River was partially indurated, possibly during subaerial exposure, before deposition of the Kings Falls. There is little relief along this contact.

The thickest section of the Gull River occurs at the axis of the fold at the far end of the quarry. Shales are rare. Fossils are uncommon but include <u>Isotelus</u>, <u>Eoleperditia fabulites</u>, small ostracods, <u>Liospira</u>, and cryptostome bryozoa. <u>Tetradium cellulosum</u> is absent. The very fossiliferous Kings Falls Limestone exhibits the somewhat typical cyclic nature of burrowed finer-grained calcarenites and calcisiltites alternating with cross-laminated, coarse-grained calcarenites and coquinites.

<u>Stop #3</u>. Small quarry (locality #C168):

At this stop, we shall (1) examine metabentonites at the base of the Kings Falls Limestone and (2) reexamine the Black River-Trenton boundary between the upper Gull River and lower Kings Falls limestones (Fig. 3). The top of the Gull River has one to two inches of relief, possibly due to scouring since bedding laminae are truncated. The shaly and argillaceous limestone interval between 6' 9" and 8' 1" appears to form a marker zone that can be traced from here northwestward to Newport (Fig. 3).

The fossiliferous Kings Falls contains cream-colored, sticky, 2 to 3 inch thick metabentonites at 12 and 29 inches. A similar 2-inch thick clay has been reported at 2 feet from Stony Creek one-half mile to the southeast (Craig, 1941; Kay, 1943, 1953). We shall evaluate the significance of these clays for time-correlation at the next stop.

<u>Stop #4</u>. City Brook (locality #C1):

The characteristics of the Gull River, Kings Falls, and Sugar River limestones will be compared with those at previous stops (Fig. 3). The Gull River Limestone lies disconformably on the quartz arenite-rich late Cambrian Little Falls Dolomite below the bridge. The lower falls is supported by the upper Gull River Limestone, ant the upper falls (Craig, 1941, fig. 5; Kay, 1953, fig. 11) is supported by the middle Kings Falls Limestone. The Rathbun member at the top of the Sugar River Limestone and the superjacent Denley Limestone will not be examined. These units are exposed in the vicinity of the upper bridge and the field beyond, but you will have to respect the NO TRESPASSING signs.

Gull River Limestone. The lower 8 feet are tan weathering, gray, quartz arenite-rich, ostracod-bearing, impure, thick-bedded, medium-textured, argillaceous limestones interbedded with a few calcareous shales up to 3 inches thick. Vertical burrows are abundant. A 3-inch thick metabentonite occurs at 6' 9" (Kay, 1943, 1953).

The upper $19\frac{1}{2}$ feet of the Gull River is composed of relatively pure, light gray weathering, dove gray, conchoidally fracturing calcilutite (sublitho graphic) and some calcisiltites. Stylolites are abundant from 11 to 16 feet. Thin shales are frequent between 13 and 16 feet, at the 18th foot, and especially between $19\frac{1}{2}$ and $21\frac{1}{2}$ feet where the limestones are very argillaceous (Fig. 3). Vertical burrows (Phytopsis) are abundant between 11 and 16 feet and in the top foot. Mudcracks occur above and below the 25th foot. An intertidal or lagoona origin seems probable for these limestones.

Kings Falls Limestone. Sediment from a coquinal calcarenite bed at the base of the Kings Falls fills some of the burrows in the highly burrow-reworked calcilutite bed at the top of the Gull River. The Kings Falls is characterized by coquinal calcarenites, as at previous localities, in contrast with the non-coquinal calcarenites of the superjacent Sugar River Limestone. Cross-laminated and pararippled beds are frequent.

At 7 feet a deep reentrant marks where a metabentonite is weathering out. Less than a mile north, at Buttermilk Creek (Fig. 3), this clay is 9 fee above the base of the Kings Falls (Kay, 1953). If this altered volcanic ash near the base of the Kings Falls between Stony Creek and Buttermilk Creek is part of a single bed, then it represents a synchronous time surface indicating that this formation is onlapping the Gull River eastward. Therefore, the base of the Kings Falls becomes progressively younger eastward, increasing the gap in time marked by the Black River-Trenton boundary in that direction. The Gull River, of course, has been dated with conodonts as Bolarian (Hasan, 1969) contrary to the correlation chart of Ordovician rocks in New York State (Fishe, 1962).

Sugar River Limestone. The contact between the Kings Falls and Suga River limestones is drawn where shale becomes more abundant. This coincides with a contact drawn where non-coquinal calcarenites become persistently abun dant and coquinal calcarenites almost disappear. The lower Sugar River at thi exposure is mainly composed of interbedded coarse-grained calcarenites and calcareous shales. These calcarenites are encrinitic and rich in bryozoa, especially cryptostomes. The shales are especially abundant in the lower 10 fet thus further accentuating lithic contrast with the upper Kings Falls below. The Sugar River Limestone contains the Cryptolithus tesselatus Zone which is characterized by C. tesselatus and the relative abundance of Prasopora. Unusually large Prasopora occur near the top, as noted at Stop #1.

Stop #5. Small quarry (locality #C105):

The purpose of this stop is to examine a black chert-bearing, subticl "calcisiltite lithofacies" of lower Trentonian age lying between $9\frac{1}{2}$ feet of Gul River and one foot of Kings Falls limestones (Fig. 3). This $3\frac{1}{4}$ foot unit is composed of massively bedded, medium gray weathering, irregularly burrowed, dak gray to black, argillaceous, and somewhat conchoidally to brittlely fracturing limestones with irregular wavy bedding surfaces separating ½ to 3 inch thick continuous and discontinuous layers. Its contacts with the Gull River below and Kings Falls above are marked by "corrasion" surfaces. Although diagnostic Trentonian macrofossils have not been identified from this exposure, conodonts (Hasan, 1969) indicate a Trentonian age for this facies and a Bolarian age for the Gull River limestones below.

The Gull River limestones are composed of light dove gray to medium gray calcilutites and calcisiltites. Fenestral fabric, horizontal laminae (algal?), limestone intraclasts, stylolites, and thin shale layers are frequent. The argillaceous marker zone of the upper Gull River Limestone (see Stops #3 and #4) occurs at the 4th foot exposed. The vertical burrow Phytopsis is common, while infrequent body fossils only occur within the upper 2½ feet, including Eoleperditia fabulites, small ostracods, Loxoplocus, strophomenid brachiopods, and Tetradium cellulosum.

Stop #6. Small hillside quarry esposures (locality #C112):

The lower Trentonian "calcisiltite lithofacies" first appears as a thin (one foot thick) but recognizable lithology at this locality (Fig. 3). It can be distinguished from the subjacent calcilutites of the Gull River and the superjacent calcarenites of the Kings Falls. Conodonts have also been studied from this exposure (Hasan, 1969). Large tabulate corals (Foerstephyllum halli) are present, as in some exposures farther northwest, but the black chert is absent. Only the two-inch thick strophomenid brachiopod-rich bed at this locality is represented at Buttermilk Creek less than a half mile south (Fig. 3). Whole Tetradium cellulosum colonies can be found in the uppermost Gull River whose argillaceous limestone marker bed is exposed between 4 and 5 feet in the lower terrace towards the road.

Stop #7. Railroad cut (locality #C102):

This exposure (Kay, 1937, fig. 28) represents the most easily reached complete section of the lower Trentonian subtidal "calcisiltite lithofacies" (Fig. 3). These 7 feet of limestones are medium gray weathering, dark gray to black, argillaceous, irregularly burrowed, brittlely fracturing calcisiltites. The medium-bedded limestones of the non-shaly lower part are separated by wavy, sometimes stylolitic, surfaces. The middle part is composed of a medium-bedded zone with black chert nodules, a few fossiliferous calcarenites, and very thin shale seams and layers. The top two feet are represented by a massive ledge containing comminuted shell material and some whole fossils. The contact with the 8 feet of exposed upper Gull River Limestone below is marked by a "corrasion" surface. The argillaceous limestone marker beds occur between one and three feet (Fig. 3).

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Field Trip 16: Mileage Log

This mileage log is designed to start at the toll booths of the Herkimer Exit (Exit #30) of the New York State Thruway. Participants of the 1969 New England Intercollegiate Geological Conference will take the Thruway from Albany to Exit #30, a drive lasting about one and a half hours. Mileage was taken from my car's odometer which reads about 3.3% too high according to the Thruway mileage markers. I believe most cars have a similar error. The hundredths of a mile are estimated, especially where turns occur in rapid succession.

This field trip visits the following quadrangles. Published geologic maps can be found in the references listed:

Little Falls 15' Quad. - Cushing (1905), Young (1943), and Fisher (1954).
*Utica 15' Quad. - Young (1943) and Kay (1953).

**InMi	CumMi	
0.0	0.0	Toll booths of the Herkimer Exit (Exit #30) of the New York State Thruway. Leave toll booths and proceed straight ahead.
0.1	0.1	Turn right, taking Route 28 North. Get in left lane.
0.2	0.3	Stop light. Turn left (north), continuing on Route 28 North.
0.25	0.55	Stop light. Junction with Route 5. Turn right (east) onto Route 5. Proceed through Herkimer (6 more stop lights) towards Little Falls.
1.75	2.3	State Police on left. Speed Limit is 50 MPH. Continue straight.
4.9	7.2	"City" limits of Little Falls.
0.9	8.1	Stop light. Proceed straight ahead. (Do not bear right.)
0.3	8.4	Stop light. Proceed straight ahead.
0.2	8.6	Stop light. Proceed straight ahead. (City Hall on right.)
0.2	8.8	Stop light. Continue straight ahead. (School on left.)
0.15	8.95	Stop light. Bear left. You are now on Routes 5 and 167 combined. Get in left lane.
0.15	9.1	Precambrian rocks on both sides of highway.
0.5	9.6	Blinking yellow light. Turn left onto Route 167 North.

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*One stop is suggested in this quadrangle.
**InMi= Incremental Mileage; CumMi= Cumulative Mileage.

1.0	10.6	Parking area on right. Fine view of Mohawk Valley.
0.2	10.8	Outcrop of Little Falls Dolomite on left side of road.
1.45	12.25	Proceed straight ahead, leaving Route 167. Pass Esso Station on your left. Now on Dockey Road.
0.15	12.4	Intersection with Bidleman Road. Proceed straight ahead.
0.4	12.8	"Y" intersection after small bridge. Bear left onto Inghams Mills Road.
0.75	13.55	Intersection with East Creek Road. Continue straight on Inghams Mills Road.
0.75	14.3	After coming down hill, continue straight onto dirt road (a dead end sign marks it). (Do <u>not</u> turn right and go onto bridge over East Canada Creek.) Note: Poor exposures of Rocklandian and Kirkfieldian (lower Trentonian) limestones and shales to left.
0.05	14.35	Turn right and cross small wooden bridge.
0.04	14.39	After crossing wooden bridge, take right fork in dirt road.
0.02	14.41	Turn left in front of building.
0.02	14.43	Turn left back onto dirt road.
0.05	14.48	Park on grass along right side of dirt road.

<u>Stop #1</u>: Walk to the right, through the grass, and proceed to the right of the wire fence, walking beneath the power line tower.

At the stone wall along the edge of the field, bear left and walk along the wire fence. <u>CAUTION</u>: Poison ivy often grows in abundance along this path.

Opposite the brick building, turn right and proceed <u>very</u> carefully over the boulders and across the creek towards the base of the outcrop. The boulders you will have to walk over to get to this exposure are sometimes unstable and move when stepped on.

0.0	14.48	Drive straight ahead on the dirt road.
0.02	14.5	Turn left onto dirt road leading from power plant.
0.05	14.55	Bear right, crossing small wooden bridge. Then, bear left.
0.05	14.6	Intersection with Inghams Mills Road. Proceed straight, uphill. (Do <u>not</u> turn left and cross bridge.)
0.8	15.4	Intersection with East Creek Road. Turn right onto East Creek Road.
2.15	17.55	Intersection with Route 167. Turn right onto Route 167, heading north.

0.4	17.95	Turn left onto Bronner Road.
0.65	18.6	Intersection with Murphy Road. Continue straight on Bronner Road
0.6	19.2	Bear right where Bronner Road turns right (intersection with Davi Road).
0.2	19.4	"Y" intersection. Bear left, continuing on Bronner Road.
1.25	20.65	Park on right side of road.
		Stop #2: Walk into small old quarry about 100 feet into field fr right side of road.
0.0	20.65	Proceed straight ahead (west).
0.15	20.8	Intersection with Burrell Road. Turn left (south)
0.1	20.9	Exposures of Shorehamian limestones on both sides of Burrell Road
0.2	21.1	Intersection with Yellow Church Road. Turn right (west) onto Yellow Church Road.
0.7	21.8	Intersection with Route 170. Proceed straight ahead. Yellow Church Road changes name to Top Notch Road.
0.7	22.5	Intersection with dirt road. Bear right, continuing on paved roa
0.45	22.95	Intersection. Continue straight ahead.
1.05	24.0	Intersection with Cole Road. Continue straight ahead. Top Notch Road changes name to Rockwell Road.
0.75	24.75	Acute angle intersection with Route 169. Proceed north on Route 69.
4.6	29.35	Crossing Stoney Creek in a relatively narrow stream valley.
0.6	29.95	Park on right side of road by entrance to quarry uphill from smal creek and outcrop of Little Falls Dolomite.
		Stop #3: Climb under barbed wire fence and walk into small old quarry about 100 feet into woods from right side of road (Route 19).
0.0	29.95	Proceed downhill (north), continuing on Route 169.
1.05	31.0	Stop light. Downtown Middleville. Proceed straight ahead onto Route 28 North.
0.7	31.7	Quarry on right is in Little Falls Dolomite.
1.1	32.8	Turn right onto newly paved road. Drive straight uphill.
0.25	33.05	"Y" intersection. Bear left, going downhill, onto Old City Road

0.15	33.2	Park on either side of road before bridge over City Brook.
		<u>Stop #4</u>: Walk across bridge and down steps on upstream side of bridge leading to stream bed.
0.0	33.2	Proceed straight ahead, crossing bridge.
0.05	33.25	(Herkimer diamonds are common in the Little Falls Dolomite downhill to the left.)
0.5	33.75	Turn sharp, acute, right onto White Creek Road.
0.75	34.5	Intersection with Elm Tree Road. <u>Side Trip A</u> begins here. See end of main road log for directions.
0.0	34.5	Turn right onto Elm Tree Road.
2.35	36.85	Intersection with Hard Scrabble Road. Turn left (north), continuing on Elm Tree Road towards town of Norway.
0.36	37.45	<u>Beware</u>: Very bad sharp right bend in road 0.2 miles ahead.
1.1	38.55	Intersection with Newport Gray Road. Proceed straight ahead up the hill. You are now driving on Newport Gray Road.
0.1	38.65	Intersection with Dairy Hill Road. Turn sharp, acute, right onto Dairy Hill Road.
0.4	39.05	"Y" intersection with dirt road. Bear left and continue on paved road.
1.45	40.5	Intersection with Black Creek Road. Continue straight on Dairy Hill Road.
1.7	42.2	Intersection with dirt road. Continue straight.
0.2	42.4	Intersection with dirt road. Bear right, continuing on paved road.
0.55	42.95	Short dirt path for a car on right. A small garbage dump on south side of path.
0.01	42.96	Turn right onto second dirt path and park.
		<u>Stop #5</u>: Walk along path (an overgrown old dirt road) for about 400 feet to small old quarry. Do <u>not</u> attempt to drive to quarry.
0.0	42.97	Back out and return (north) the way you came.
0.53	43.5	Intersection with dirt road. Bear left, continuing on paved road. Continue straight ahead through the next two intersections.
3.35	46.85	Bear right and continue on paved road.

0.45	47.2	Turn sharp, acute, left onto Newport Gray Road
0.2	47.4	Continue straight downhill on Elm Tree Road.
1.7	49.1	Turn right and continue on Elm Tree Road.
2.25	51.35	Turn left onto White Creek Road.
0.8	52.15	Bear right at intersection, continuing on White Creek Road.
0.7	52.85	Intersection with Route 28. <u>Side Trip B</u> begins here. See end of main road log for directions.
0.0	52.85	Turn left onto Route 28 South, heading towards Middleville.
1.3	54.15	Herkimer diamonds common along exposure of Little Falls Dolomite on the left side of road.
1.2	55.35	Stop light. Downtown Middleville. Turn right and cross bridge over West Canada Creek, thus continuing on Route 28 South.
0.2	55.55	Bear left at fork in road, continuing south on Route 28.
8.15	63.9	Turn right (west) onto Route 5 (Routes 5 and 28 combine here for a short distance). Proceed into Herkimer.
0.7	64.6	Stop light. Intersection of Routes 28 and 5. Turn left (south) onto 28 South.
0.2	64.8	Stop light. Turn right, continuing on Route 28 South. Get into left lane.
0.25	65.05	Turn left onto Thruway entrance.
0.1	65.15	Toll booths for Exit #30 (Herkimer Exit) of the New York State Thruway. Return to Albany via Thruway.

<u>Side Trip A</u>:

0.0	0.0	Continue straight on White Creek Road.
0.15	0.15	Park on right side of road by intermittent creek.
		<u>Stop #6</u>: Climb under barbed wire fence on right side of road. Walk about 300 feet into field until you reach the second set of outcropping ledges.
0.0	0.15	Drive straight ahead.
0.35	0.5	Intersection with Woodchuck Hill Road. Turn around and go back on White Creek Road the way you came.

0.45	0.95	Intersection with Elm Tree Road. Turn left and continue with main part of field trip.

Side Trip B:

0.0	0.0	Intersection of White Creek Road and Route 28. Turn right onto Route 28 North, heading towards Newport.
2.05	2.05	Stop light. Newport, N. Y. Turn left onto Bridge Street.
0.15	2.2	Cross bridge over West Canada Creek.
0.1	2.3	Intersection. Turn right onto Newport Road.
0.8	3.1	Large old quarry to the right contains a good exposure of Gull River Formation (Bolarian Series).
1.4	4.5	Park on right side of road next to barbed wire gate in barbed wire fence. Poorly exposed Kings Falls Limestone (Kirkfieldian) about 100 feet ahead on left side of road.
		Stop #7: Walk across the field along right side of road. Beware of the bull! Head towards the small clump of trees at the opposite side of the field where you can climb beneath the fence and down to the railroad tracks.
0.0	4.5	Proceed straight ahead.
0.15	4.65	Intersection with dirt road on your left. Turn around and return (south) to Bridge Street.
2.35	7.0	Turn left onto Bridge Street.
0.25	7.25	Turn right (south) onto Route 28 South (Main Street).
2.05	9.3	End of Side Trip B. Return to New York State Thruway from this intersection with White Creek Road according to main road log.

Trip 17

APPLIED GEOLOGY IN THE CENTRAL HUDSON VALLEY

Road Log

by

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Topographic maps (1:24 000) for the trip include:

Albany
Delmar
Ravena
Hudson North
Alcove
Leeds
Cementon
Saugerties
Kingston East

The chief purpose of this trip is to lay out the use of geology in the cement industry of the Hudson Valley, with the main emphasis on the quarry of the Hudson Cement Company at East Kingston. This is geologically the most complex of the New York cement operations and a program of continuing geologic control has been in effect since 1961. The road log will indicate the stratigraphy passed as the bus travels the New York State Thruway, although no stops can be made. Economic use of the various stratigraphic units will be emphasized.

Miles

0.0 Leave Thruway Hyatt House; turn right on Washington Avenue. Building complex on left is the new Albany Campus of the State University of New York, designed by Edward Durell Stone. Abundant use has been made of white cement (not a product of the Hudson Valley) and architectural aggregate, both in pre-cast panels and in cast-in-place concrete. Test borings at the campus site showed over 200 feet of clay.

Miles

0.9 Turn right on Fuller Road and follow signs to Thruway. The pine covered hills are sand dunes above the glacial Lake Albany Plain. Some dunes have crude barchan forms with the barbs to the west.

2.1 Enter New York Thruway. Bear left toward New York City. In the few short years since the Lake Albany beds were laid down, the area has been deeply dissected. The thick clays (up to 300 feet) of the lake bed are unstable in some of these valleys (or "kills," to use the Dutch word seen so often in this area) and landslides have been a problem. To the right can be seen the Helderberg Escarpment, the low ridge of Devonian limestone, and behind it the Catskill Mountains.

12.2 (138.6 on Thruway markers) East side. Road cut with Ordovician shales (Normanskill formation).

15.1 (135.7 on Thruway markers) Road cut with interbedded shale and graywacke of Normanskill formation.

Occasionally, operations can be seen along the Thruway or the Northway where Albany Molding Sand is removed. This is a dying business which once flourished from Kingston to Glens Falls. The sod is carefully removed to expose a thin (2 feet or so) layer of quartz sand bonded by clay. After removal of the commercial molding sand, the sod is replaced.

18.7 (132.0 on Thruway markers) To the west can be seen the plant of the Atlantic Cement Company. Below the bridge is the covered conveyor belt which carries the finished product from the plant site to the dock on the Hudson River. Geologists were involved in the search for suitable stone, in the selection of the plant site and the conveyor route, and in choosing the optimum location for the loading dock so as to meet engineering requirements without positioning over expensive brick clay reserves. The initial investment here was $44,000,000. Production comes from the Manlius, Coeymans, Kalkberg, Becraft and Esopus formations along the Helderberg escarpment west of the plant. The beds lie in gently-dipping folds cut by flat thrust faults which offset the beds as much as 10 feet.

21.8 (128.9 on Thruway markers) Road cut in Normanskill graywacke. To the west is the Hudson River Concrete Products plant. The raw materials for the concrete blocks are crushed Kalkberg limestone from a quarry about one mile to the southwest, local sand and cement Along the Hudson River, a string of brick plants grew up utilizing the Lake Albany clays. At one time there were 129 producers from Westchester County to Glens Falls. Now only a handful survive.

Miles

23.8 (126.9 on Thruway markers) New Baltimore Service Area. Geologists for the New York State Department of Transportation were given the assignment recently of augmenting the water supply, not an easy job in an area where the surface materials are deep clays and the bedrock is shale. A suitable well was found on the east side of the highway and a tunnel had to be constructed under the road bed.

27.4 (123.7 on Thruway markers) As the road rises to the south, we will pass from the Normanskill to the Helderberg group.

29.1 (122.0 on Thruway markers) East side: Lower Hannacroix member of Kalkberg formation; note black chert bands. West side: long cut contains all units from Manlius through New Scotland in a very complex structure.

30.5 (120.6 on Thruway markers) Kalkberg formation .

30.8 (120.3 on Thurway markers) New Scotland formation.

31.0 (120.1 on Thruway markers) Becraft formation.

31.9 (119.2 on Thruway markers) Esopus formation. Note the difference in weathering response between the clean Becraft limestone road cuts and the highly fragmented Esopus shale cuts. The latter indicates a high sensitivity to wet-dry or freeze-thaw alternations. A breakdown of the fine material at the base of a cut is an almost certain indication that the rock is unsuitable for use as crushed stone.

32.2 (118.9 on Thruway markers) Becraft formation, steeply dipping.

32.7 (118.4 on Thruway markers) New Scotland through Becraft.

33.0 (118.1 on Thruway markers) Esopus formation.

33.4 (117.7 on Thruway markers) Schoharie and Esopus formations.

34.1 (117.0 on Thruway markers) Onondaga formation. Note chert layers.

34.7 (116.4 on Thruway markers) Highly fractured Becraft. These calcite-healed fractures suggest nearness to a thrust fault.

35.1 (116.0 on Thruway markers) Note long, north-south lake on right. Structure and stratigraphy suggest the lake marks the position of a thrust fault, a common occurrence in this area.

Miles

37.3 (113.8 on Thruway markers) East side. Note New Scotland faulted onto New Scotland.

37.4 (113.7 on Thruway markers) Becraft above New Scotland. The bridge crosses Austin Glen, the type locality of the Austin Glen member of the Normanskill formation.

38.2 (112.9 on Thruway markers) Esopus below Schoharie.

39.2 (111.9 on Thruway markers) Schoharie. The cut in the median strip is synclinal.

39.6 (111.5 on Thruway markers) Esopus-Schoharie-Onondaga.

41.0 (110.1 on Thruway markers) Schoharie-Onondaga.

41.3 (109.8 on Thruway markers) Cut east of Thruway. Esopus with lowest portion high in bedded chert.

42.0 (109.1 on Thruway markers) Esopus below, Schoharie above with broad transition zone.

43.1 (108.0 on Thruway markers) To east of Thruway lie three cement plants clustered in the beds of the Helderberg group. South to north they are: The Alpha Portland Cement Company; The Lehigh Portland Cement Company; and the Marquette Cement Manufacturing Company

47.3 (103.8 on Thruway markers) Port Ewen formation above Becraft. Note the large blocks that break loose from the face of the road cut. These are known to engineering geologists as wedge failures. These bedding plane and joint intersection problems have largely been eliminated by pre-split blasting of the final face. We will see this at a later stop.

48.8 (102.5 on Thruway markers) Prepare to leave Thruway.

49.5 (101.6 on Thruway markers) Leave Thruway at Exit 20.

49.6 Turn left (south) on Route 32.

49.8 Turn left (east) on Routes 32 and 212.

50.1 Turn right at sign pointing to Mt. Marion. Follow south on county road parallel to Thruway.

50.5 Onondaga formation.

Miles

51.6 Note historic house on east side. Many stone houses can be seen in this area. Formations used for such construction include Becraft, Onondaga and New Scotland as well as middle and upper Devonian graywackes from the Catskills.

52.5 East side. Plant of Hudson Lightweight Aggregate Corporation. Esopus siltstone is quarried in the ridge to the east. Note the twin rotary kiln and cascading slag. Another lightweight aggregate quarry will be visited later in the day.

53.7 Turn left (east) at cross road.

54.1 Road cuts in Esopus. Note the cedar growth on the Esopus. This gives a rough suggestion of underlying rock type where outcrops are lacking.

55.1 Turn right (south) on Route 9W.

55.8 East side. Glenerie limestone.

56.3 West side. Glenerie Falls of Esopus Creek.

56.9 East side. Dip slope of Glenerie limestone.

57.2 East side. Note mushroom plant. The mushroom growing industry once used the abandoned natural cement underground mines because of their constant temperature, darkness and low initial construction expense. Current production is carried out in specially constructed buildings which enable lower labor costs. One former cement mine (near Rosendale) is used for storage of records, sent chiefly from various company offices in New York City.

59.6 Turn right at the second of the two right turns of the clover leaf toward Rhinecliff Bridge on Route 199 going east.

60.1 Long road cut. First rock is Schoharie, followed by Esopus. Note that all of these cuts have the typical smooth face and closely-spaced drill holes of the pre-split blasting method. In the campaign to adapt the method to New York conditions a prominent role was played by New York Department of Transportation geologists.

60.4 Note long north-south lakes, probably developed along eroded thrust faults.

60.6 Port Ewen, Alsen, and Becraft formations in an anticline.

Miles

60.8 Sequence: New Scotland, Kalkberg, Coeymans, Manlius. The scarp face on the east marks the end of the Helderberg beds.

61.1 Turn right toward Route 32.

61.2 Turn left (south) on Route 32.

61.5 West side. Contact of Rondout above the Normanskill. This is the major unconformity of the area and is probably a detachment fault.

61.6 STOP 1. (71.50 N - 59.82 E; Kingston East Quadrangle) Road cut on Route 32. 2,000 feet south of intersection of Routes 32 and 199 Time enough will be spent here for familiarization with the units from the Normanskill through the Kalkberg, so that they may be recognized from a distance in the quarry. Note well-marked east-dipping thrust faults and poorly-marked west-dipping thrust faults. Stromatoporids are uncommonly abundant in the Manlius formation.

62.2 West side. Note old vertical kiln used to burn Manlius for lime.

62.3 Turn left on road to East Kingston.

62.4 On west can be seen portals and airways of an old natural cement mine. Here the Whiteport member of the Rondout was mined. This is the northernmost of the many natural cement mines that flourished from here to beyond Rosendale. Of the dozens of onetime operations only the Century Cement Company still operates in the belt. Both the Whiteport and the Rosendale members were utilized by the various companies.

63.0 Turn left on John Street.

63.4 Approaching the Hudson Cement plant. On the left is the site of an old brick plant which has been completely removed to make way for the cement facility. The red brick barns on the west of the road are all that remain; they are typical examples of the soft mud brick made in the valley.

63.8 STOP 2. (70.92 N - 60.00 E) Offices of the Hudson Cement Corporation, River Street, Kingston, New York. This is the most complex quarry in the valley, both because of geologic complications and because of the multiple products. Consequently, geologic control has been vital since the early days of operation. The highly folded (even overturned) beds and the abundant thrust faults, many of which are folded, combine to present the operators with ever-changing conditions. To the geologist they present an unparalleled opportunity to study thin-skin tectonics.

Stratigraphy

All units from the Normanskill through the Esopus can be seen, and of these only the Normanskill, the Connelly and the Glenerie are not currently utilized. In general, crushed stone is produced from the Broncks Lake and the New Scotland units. Cement is produced from various combinations of Manlius, Coeymans, Hannacroix, Becraft, and Alsen units. Both products come from the same working faces, so there is a shovel control problem. At times the products are blasted separately and at times together, with the separation made by control of shovel limits within the pile of broken stone. Wherever possible, the rocks are worked so that the faces are normal to the strike. Because chemical analyses are known for each distinctive layer in the quarry, it is possible to know the analysis of a blast within working limits even though the beds may be inclined or faulted.

Structure

In portions of the property, the rock lies in as many as three thrust sheets, with varying dips in each. It is a rare face that does not have at least one fault, generally an east dipping thrust fault, but possibly a west dipping thrust fault or an expanded crest that is criss crossed by obscure flat faults. Although the style of folding is flexural slip, a suggestion of "expanded crests" can be seen in some folds. Folded thrusts are seen, as well as all orders of bedding plane faults. The relation of beds to faults changes as the face is carried in the strike direction, and a constant watch must be kept on the analysis of rock being supplied to the cement plant. At Hudson the blast hole cuttings are not analyzed; instead the control is maintained by strike-quarrying and by matching production analyses with the sources. By having several faces available for shovelling, it is possible to change the mill feed by shifting to a face with higher carbonate, silica, or alumina analysis, whichever is needed.

Production

A special problem at Hudson arises from the fact that crushed stone, high analysis cement rock, and low analysis cement rock all must come from the same faces. They do not, however, occur in those faces in the same ratio as the plant requirements. This creates scheduling difficulties and in effect requires more working faces and longer haulages.

The associated lightweight aggregate plant utilizes Esopus siltstone, which comes from a separate quarry on the property.

Simply summarized, the geologist's job is to blend the geologic facts with the production requirements. He must know the quarrying methods and equipment, something of cement chemistry and production, and be familiar with the variations in the cement raw materials that do not come from his quarry. In this plant the latter are the pyrite "cinder," the coal, and gypsum. Above all, he must be able to communicate the geologic realities to the production staff.

END OF TRIP. Return to Thruway Hyatt House by way of New York Thruway.

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GENERALIZED STRATIGRAPHY OF THE MID-HUDSON VALLEY

With notes on economic uses

	Thickness	Lithology	Use
ONONDAGA FORMATION		Limestone with chert	
SCHOHARIE FORMATION		Limy shale	
ESOPUS FORMATION	150+	Shale and siltstone	Lightweight aggregate
GLENERIE FORMATION	25 ?	Limestone	Crushed stone
CONNELLY FORMATION	10	Conglomerate, sandstone	Crushed stone
PORT EWEN FORMATION	100	Argillaceous limestone	Crushed stone
ALSEN FORMATION	20	Limestone	Cement rock
BECRAFT FORMATION	50	Limestone	Cement rock
NEW SCOTLAND FORMATION	110	Argillaceous limestone	Crushed stone
KALKBERG FORMATION			
Broncks Lake Member			
Upper Broncks Lake	27	Silty limestone	Crushed stone
Lower Broncks Lake	15	Silty limestone	Crushed stone
Hannacroix Member			
Upper Hannacroix	11	Silty limestone	Crushed stone
Lower Hannacroix	17	Cherty limestone	Crushed stone
COEYMANS FORMATION			
Ravena Member	17	Limestone	High analysis cement rock
MANLIUS FORMATION			
Thacher Member			
M6	6	Magnesian limestone	"
M5	2	Magnesian limestone	"
M4	7	Limestone	"
M3	25	Limestone	"
M2	9	Magnesian limestone	
M1	5	Limestone	
RONDOUT FORMATION			
Whiteport Member	8	Dolomitic limestone	Natural cement
Glasco Member	10	Limestone	
Rosendale Member	5	Dolomitic limestone	Natural cement
Wilbur Member	14	Limestone	

Some aspects of conglomerates in the Taconics, Cossayuna area, New York

by

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I finished my mapping in the Cossayuna area a decade ago. When I returned this summer to look around for suitable stops for this guidebook, I found several aspects of the Taconic situation changed. When I started mapping, even the stratigraphic sequence was not certain. Certainly the status of the Taconic klippe was in doubt. Now, however, the sequence is well defined (Shumaker, 1967), and few doubt the existence of the klippe (Cady, 1968). Only the southern continuation of the allochthonous rocks remains to be established (Platt, 1962), although some progress is being made in this direction (Carswell, and others, 1968). Thus a tour to establish the stratigraphic sequence in this area seems uninteresting, in view of the remarkable progress made by Dale (1899).

My purpose in visiting these several outcrops in the Cossayuna area is to suggest a relation between the sedimentary and structural features and the paleotectonic setting during their formation. Two types of conglomeratic rocks will be examined, one within the allochthonous Taconic sequence, and one in the autochthonous shale west of the Taconic fault.

The Taconic sequence was deposited between a carbonate-quartzite sequence on the west flank of what is now the Green Mountains and a shaly and volcanic sequence on the east. The carbonate-quartzite sequence is approximately 5000 feet thick near the original home of the Taconic sequence (Thompson, 1967, p. 69), and the eugeosynclinal rocks on the other side are thicker (Skehan, 1961, p. 65-96). Yet the Taconic sequence is thinner than either of these, probably on the order of 3000 feet in the Cossayuna area. Why is it thinner?

Although various unconformities have been reported in the Taconic sequence, work of the last decade or so (Berry, 1959;Bird and Rasetti, 1968; Theokritoff, 1959) has been in the direction of removing the need for them and, in fact, encourages me to believe that the Taconic sequence is a comprehensive one. It extends from the Lower Cambrian, perhaps the Eocambrian, to the Middle Ordovician. As the number of fossil gaps was progressively reduced, the concept of a starved basin of deposition was considered. Another interpretation to emerge was that of a geanticline separating a basin of shale from a basin of carbonate deposition.

The interpretation that I favor is that the Taconic sequence was deposited on a slope sufficiently gentle and deep to permit the settling of clay-size material but sufficiently inclined to cause progressive sliding and slumping and general attenuation of the section. Evidence includes progressive eastward reduction in size of carbonate pebbles and in total carbonate, orientation and type of soft-sediment deformation features, western provenance of the carbonate in Lower Cambrian and Lower Ordovician rocks (stops 2 and 3) and of the quartz grains in the Upper Cambrian unit (stop 1). Thus during the Cambrian and Early Ordovician there was an eastward-facing slope between the carbonate bank and the shale area. This idea may have been put forward first by Bailey, and others (1928), and it seems supported on uniformitarian grounds by such recent data as that of Rona and Clay (1967) and Dickey, and others (1968), and on experimental grounds by Van der Knaap and Eijpe (1968).

The paleogeography during the late Middle Ordovician changed to an entirely different picture. The carbonate bank was covered by mud and then the Taconic slide came onto the mud. The mud and the Taconic slide came from the east, along with quite a variety of bits and pieces (Hawley, 1957; Ruedemann, 1930, p. 104-117; Zen, 1959, p. 8-9). Stops 5 and 6 examine some of these pieces.

Road log. Mileage cumulative from Saratoga turn-off, exit 14 on Interstate 87.

From motel go to Interstate 87 "Northway". Go on it approximately 30 miles to Saratoga. At 22 miles and thereafter roadcuts show the Middle Ordovician autochthonous shale. Take exit 14.

0.0 Keep right, joining NY Rte 9P north toward NY Rte 29. Even though the sign says you are going north on 9P and you know you are going west over the Interstate highway, this is the right way to go east to Schuylerville.

0.3 Turn right "to 29"

1.2 Stop sign. Turn right onto NY Rte 29 toward Schuylerville.

1.4 Pass under Interstate 87; now you are going east. Continue east on NY Rte 29 toward Schuylerville.

8.3 Pass through Grangerville as quickly as you can--observing the speed limit!

9.6 NY Rte 338 turns off to the right. Here note the view to the east. The near ground is underlain by autochthonous upper Middle Ordovician shale and Pleistocene deposits. The middle distance ridge line is underlain by Lower Cambrian shale and siltstone, part of the Taconic sequence which lies on the autochthonous shale. The high hills on the skyline are made of the high Taconic sequence, lithologically similar to the low Taconic sequence, but unfossiliferous because of higher metamorphic grade and a higher intensity of deformation.
Continue east on NY Rte 29.

10.3 Enter Schuylerville. The name reminds us of early Dutch influence here.

10.6 Meet NY Rte 4. Turn right onto "Broadway" keeping on Rte 29.

11.0 1st red light. Turn left with Rte 29 onto Ferry Street.

11.3 Hudson River, another Dutch name. We also here leave Saratoga County and enter Washington County, both names of historical interest.
Continue east on Rte 29.

13.9 Join NY Rte 40. Bear left.

14.2 Bridge over Batten Kill. To the left, under the trees, Lower Ordovician carbonate rests on Middle Ordovician shale, with the contact dipping east.

14.3 Bear half left. This back road has scattered along it several exposures of Bald Mountain Limestone, a unit with a long and involved history. The name, itself, has been in the literature for 125 years (Mather, 1843; Emmons, 1846). For further discussion of this unit, see Rodgers, this guidebook, stop 1.

15.9 Stop sign. Turn right.

16.2 Large dead elm is on the Taconic thrust. Exposed in the old orchard to the west is the parautochthonous Lower Ordovician Bald Mountain Limestone (not completely autochthonous because it is known to lie on the Martinsburg Shale just to the south at Middle Falls, yet not really allochthonous in the sense of far traveled because it is of the same general carbonate facies as rocks of equivalent age in the vicinity, thus parautochthonous or approximately in place). Immediately to the east is the Lower Cambrian Bomoseen Formation, seen in field exposures. The fault crosses this road three times in less than half a mile; thus here it is quite flat.

16.6 Stop sign. Intersect NY Rte 40. Turn left (north). View back into Hudson Valley from Taconic scarp.

16.7 Road cuts in Lower Cambrian Bomoseen.

17.0 On right, 9-foot dug well never ran dry until the road was widened. Then it never ran again.

17.2 Pencil-shaped pieces of Lower Cambrian Bull Formation. Various field exposures of this unit follow.

18.4 On right are exposures of Lower Cambrian West Castleton Formation. A fault, here putting the Bull Fm. onto the West Castleton, lies close to the road for a few miles.

21.5 South Argyle. Turn right toward Cossayuna Lake.

22.8 STOP 1. (Park on north side of road)

The roadcut shows the Upper Cambrian Hatch Hill Formation dipping steeply east. Although the outcrop is mostly resistant sandstone and siltstone, the formation is largely black shale, and approximately 350 feet thick. The distinctive well-rounded quartz grains in a dolomitic matrix, the coarse, frosted grains, the rusty weathered surface with quartz veins standing out in relief (not seen here) provide ready recognition of field outcrops.

The point of interest here is conglomerate of sandstone and siltstone similar to the matrix. Also note the thinning and thickening of a few beds of siltstone between beds of finer black shale, and the wispy and discontinuous nature of the finer laminae. Near the western end of the outcrop a miniature nappe showing eastward movement contrasts with the regional westward overturning in the Taconic sequence. These four features taken together with reasonable paleogeographic reconstructions and the probable source of the coarse, frosted sand grains in the Potsdam Sandstone to the west indicate an eastward slope on which this formation was deposited. Cross-lamination also shows eastward movement of material. This outcrop is near the western edge of the Taconic klippe; thus it does not indicate the length of the paleoslope. A similar conglomerate in this unit is exposed 0.4 miles north of West Hebron, on the eastern edge of the Cossayuna quadrangle, a few miles across strike.

The Middle Ordovician deformation is also shown in this outcrop by the little folds accompanied by gash veins, and by a nearly flat fault cutting the two thick sandstone beds at the eastern end of the outcrop.

Continue east.

23.3 STOP 2. (Park on north side of road. Please do not block driveway)

This pause at the Lower Ordovician Schaghticoke or Lower Poultney will be a short one because it is not very convincing for my story. The unit is upside down, though this is not easy to demonstrate here. Note the fine grain of both the carbonate and the shale, surely an indication of quiet water, yet some grading and some cross-lamination and a few tiny stylolites can be found. There is also some irregular thinning of carbonate beds. Because the thinning is irregular and irregularly distributed in the outcrop, I believe it was caused penecontemporaneously with deposition. Elsewhere in the unit, as far east as the Pawlet quadrangle, the limy layers have been completely pulled apart and appear as conglomerates, in some places with angular pieces (Shumaker, 1967, p. 26).

Continue east.

24.2 Excellent roadcut in pseudo-conglomerate of the Lower Cambrian West Castleton Formation. The way the soft, limy blebs compacted and fitted together can be seen here.

24.7 STOP 3. (Park on left before intersection)

At the southwest end of Cossayuna Lake, the roadcut is on the west, overturned, limb of a large anticline mapped from the northeast corner of the Cossayuna quadrangle south-southwest to the southern edge of the quadrangle and cored with 1000 feet of Bomoseen Grit of Ruedemann (1914, p. 69), part of the Bull Fm. of Zen (1959). This unusually good exposure is at the same stratigraphic level as the roadcut two thirds of a mile west, on the other limb of a small syncline. Similar but thinner conglomerates are exposed near the Vermont line in the eastern part of the Salem quadrangle on the north side of Dry Creek and at Buttermilk Falls Brook, where I have been mapping.

Features suggesting soft-sediment slumping include the limy blebs, flakes of dark shale and dark siltstone, puckering of several thin layers of siltstone with a sense of motion to the east (outcrop overturned), curling of the shale flakes, the fitting together of the limy blebs--rather like pillow lavas, and the lack of some coarse matrix such as one might expect from violent turbidity currents. The impression I receive is that quiet deposition of limy mud and clay mud and some silt was disturbed by occasional gentle slipping down a gentle slope. Perhaps the slipping was more or less constant, so that the beds were continuously being attenuated. This might be invisible in clays with no coherence, but the calcium carbonate mud may have rapidly become sticky enough to retain some coherence (Hathaway and Robertson, 1960) when pulled by the force of gravity down a very slight eastward slope. Thus there are more or less even rows of blebs in undeformed shale. If this were tectonic boudinage, calcite veins might be found between the pulled-apart blebs. There is, in fact, abundant calcite veining, but preference for limy layers is not evident.

Turn left (north) up the west side of the lake.

25.9 Several outcrops of Lower Cambrian Bomoseen.

27.2 Turn left (west) at Billiken.

28.3 From crest of hill note the view in each direction. To the east the high Taconics--to the west the Hudson Valley and the Adirondacks.

28.6 At base of hill join County road 47. Continue west.

28.8 Corner between Dutchtown Rd. and Street Rd. This name seems redundant; actually it refers to The Street of the Argyle Patent, 1764. Continue west.

29.4 Bear left.

29.7 Bear right onto bumpy road.

29.8 On left good exposures of Bomoseen.

30.2 STOP 4. (Park across intersection. Snakes are fairly common around here.)

These interbedded limy and silty beds are Lower Cambrian because Dale (1899, plate 13) reported Olenellus from two places in this field. A search around the pasture will show the limy layers are pulled apart in places. It is not certain here whether the pulling apart was during Early Cambrian deposition or during late Middle Ordovician deformation. For example, one might propose that because the calcite veins go through between the calcite-rich pods in some cases (see chevron folds), the boudinage and the veining follow the same stress pattern and are of the same cloth. Or the places between pods may have been weak enough to localize the formation of veins. In fact, it is not certain everywhere at this stop what feature or features cause the weathering pits to have their present shape.

Continue west, down hill across the Taconic thrust.

30.9 Crossroads with NY Rte 40. Turn right toward Argyle.

31.0 Roadcuts on left are in the autochthonous shale but contain samples of Taconic rock types. In fact, we have here a different kind of conglomerate from that seen in the first few stops, for the source is to the east. Thus the slope faced west by the time this conglomerate formed (late Middle Ordovician).

31.7 Center of Argyle. Lunch stop in Fire House in case of bad weather. Gas, etc.

Continue north, now on NY Rte 197.

32.5 Keep straight ahead. Then turn right.

33.1 Ledge of typical Martinsburg Shale. I use that name because it really does fit reasonably well. The name Snake

Hill has been abandoned following Berry's paper (1963), and this fairly fine-grained unit is not typical of the Austin Glen Graywacke. Note the prominent cleavage and the difficulty of seeing bedding. About 200 feet along the ledge, the rock suggests pieces of shale in shale.

33.7 Intersection with NY Rte 40. Turn left.

34.1 Turn right.

34.6 STOP 5.

Field to right has ledges of conglomerate in the Martinsburg Shale. But the conglomerate is carbonate pebbles, and there is no shale matrix. Few if any of the pebbles are definitely attributable to the Taconic sequence. The rock is identical to outcrops near Bald Mountain called the Rysedorph Hill Conglomerate by Ruedemann (1914), and it seems to present a paradox in provenance because it is immediately west of and beneath the Taconic klippe. The Taconic sequence covered the carbonates in the area from which these pebbles came at the time they came!? Such lenses of carbonate conglomerate seem to be confined to a narrow zone close in front of the klippe, so one might suggest that they were entrained by the klippe. But entrained how, and from where, if the autochthonous shale blanketed the carbonate sequence? Alternatively, the conglomerate may have been formed earlier, near the top of the slope on which the Taconic sequence was deposited; when the slope was reversed, the conglomerate would have moved as single big blocks. No doubt there are other possible explanations, but the lack of shaly matrix must be accounted for.

Continue east.

34.8 Intersection. Turn right approximately on the Taconic thrust, for the hill to the east is Lower Cambrian Bull Fm. and the lowlands to the west are Martinsburg Shale, here late Middle Ordovician.

35.4 Road joins from left.

35.5 Crossroad. Turn right.

36.6 At NY Rte 40, turn left (south) into Argyle.

37.5 Turn left with NY Rte 40.

38.3 Crossroad. Turn right (west).

38.7 STOP 6. (Turn into driveway, but please do not block it)

West of the patch of trees are several exposures of the Martinsburg Shale and of pieces of various rocks in it. Several resemble rocks in the Taconic sequence. This stop is right on strike with the last one. Several miles to the south is Bald Mountain.

As I have a long drive back to Washington, this is the end of the trip. To return to Albany, return east to NY Rte 40. Go south to NY Rte 29. Go west to Interstate 87, and south.

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Trip 19

BEDDING THRUSTS AND OTHER STRUCTURAL FEATURES IN CROSS SECTION THROUGH "LITTLE MOUNTAINS" ALONG CATSKILL CREEK, (AUSTIN GLEN AND LEEDS GORGE), WEST OF CATSKILL, NEW YORK

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INTRODUCTION

The valley of Catskill Creek, a tributary of the Hudson River, cuts a transverse section through the numerous folds and faults of the Hudson Valley's miniature Appalachian Valley and Ridge Province, which W. M. Davis (1882) termed "The Little Mountains east of the Catskills." This transverse section extends from Jefferson Heights, Catskill, on the southeast (NE part of the Cementon 7 1/2-minute quadrangle; the NW quarter of the Catskill 15-minute quadrangle), to Leeds, on the northwest (SE part of Leeds 7 1/2-minute quadrangle; the SW quarter of the Coxsackie 15-minute quadrangle). The exposures along Catskill Creek in Austin Glen and Leeds Gorge illustrate many facets of Appalachian stratigraphy, sedimentology, and structural geology.

This excursion is intended to display the sedimentary features of the units, particularly those of the Austin Glen Member of the Normanskill Formation, and the geologic structure. The "Little Mountains" are a classic area. Their strata and fossils are continually being renamed to keep them "up to date" as revisions and detailed studies occur, but in general no correspondingly extensive up-to-date geologic mapping has been carried out to keep their structural interpretations abreast of modern knowledge and of the new 7 1/2-minute topographic quadrangle maps. The present writer's reasons for conducting this trip and for preparing this guidebook are: (1) Austin Glen and Leeds Gorge are among the writer's favorite places, both scenically and geologically; (2) the strata are well exposed and illustrate the contrasting products of deposition, in waters of various depths in the Paleozoic seas, of noncalcareous flysch on the one hand and of calcareous nonflysch strata on the other; (3) the information, recently exposed in roadcuts, about the geometry of the bedding thrusts suggests that certain currently fashionable ideas favoring the "Acadian" age of the deformation of the Silurian and Devonian strata in the Hudson Valley should be discarded in favor of the earlier, now out-of-fashion interpretation that the deformation belongs with the terminal

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Appalachian orogeny (dated only within the interval mid-Permian to Late Triassic); and (4) the recent study by E. A. Babcock 5th has contributed new information on the relationships among folds, cleavage, and fractures.

Although the excursion will not attempt to examine all the evidence on the subject, this guidebook article will discuss the "Little Mountains" as an example of the dying out of the Appalachian folds, both to the north and to the west; of the effects of the competency of the strata on the sizes of folds; and of the structural effects of thickness of sedimentary cover above the "rigid" basement. In addition, this article will comment on the Ruedemann-Ulrich relationship with regard to the "Normanskill problem," which for more than half a century has served as a complicating factor in the interpretation of Hudson Valley stratigraphy and structures.

GEOLOGIC FORMATIONS

The geologic formations in Austin Glen (Table 1) are entirely sedimentary and range in age from medial Ordovician (Austin Glen Member of Normanskill Formation) to medial Devonian (Onondaga Formation). These geologic formations fall naturally into 2 groups: (1) an older, noncalcareous, graded graywacke-siltstone complex of Mid-Ordovician age, whose chief fossils are graptolites, and whose strata have not been subdivided but are all included within the Austin Glen Member of the Normanskill Formation; and (2) a Silurian-Devonian succession, including many kinds of carbonate rocks and siltstones, containing abundant brachiopods, and whose strata have been extensively subdivided into numerous formations and members (Table 1).

Although in Austin Glen a noteworthy hiatus exists at the contact between the Upper Silurian Rondout Dolostone and the underlying Middle Ordovician Austin Glen Member of the Normanskill Formation, the strata of these two units in valleyside exposures are parallel. In Catskill Creek at low water, however, Chadwick (1913) reported that these formations are in angular unconformity. Only a few kilometers from Austin Glen the Upper Silurian strata likewise rest with profound angular unconformity on the Ordovician or older rocks (Schuchert and Longwell 1932; Chadwick and Kay, 1933).

Stratigraphy

This section on stratigraphy will concern the names of the formations, their thicknesses, and the general characteristics of the rocks. The details of sedimentary structures and interpretations of environment of origin are contained under the heading of sedimentology.

<u>Austin Glen Member of Normanskill Formation</u>

Austin Glen is the type locality for the Austin Glen member of the Normanskill Formation (Ruedemann, 1942). Although these strata are well

Table 1. Formations of the "Little Mountains" near Catskill Creek, Austin Glen to Leeds Gorge, based on published sources and original observations. (Onondaga, after Oliver, 1962; Schoharie, after Johnsen and Southard, 1962; Port Ewen through Rondout, after Rickard, 1962)

Formation	Member		Thickness (ft)	Characteristics
Onondaga	Moorehouse	Middle unit	25+	Top not exposed; Ls., fine gr, med dk gray; abundant dark gray chert; corals & bryozoans common; also brachiopods and other fossils.
		Lower unit	12	Limestone, medium, light to med gray; no chert; corals common; other fossils present.
	Nedrow		4	Limestone, med. gr; light to med gray; a few chert nodules; platyceratids, corals, and brachiopods; transitional to Moorehouse
			39	Limestone, med to cs. and fn gr; light to med gray; abundant light to med gray chert; platyceratids, corals, and brachiopods.
	Edgecliff		32	Limestone, med and cs-gr; light to med and light gray; abundant light to med. gray chert; large crinoid columnals common; also corals and brachiopods.
			3	Limestone, fine gr; med gray; no chert; fragmental fossils.
	Transitional interval		1 1/2	Limestone and gritty limestone alternating.
Schoharie	Saugerties		16	Limestones; interbedded light gray-weathering and yellowish-brown weathering muddy limestone.

Schoharie	Aquetuck	40	Mudstone, calcareous, dark gray, weathering yellowish-brown; uppermost 3 ft. sandy and glauconitic; includes indistinctly set off, light-weathering muddy limestone beds; abundant indistinct layers of dark-weathering rock and layers of small chert nodules.
	Carlisle Center	13	Mudstone, calcareous, dark gray, weathering yellowish-brown and yellowish-gray; includes interbeds that are slightly limier than the mudstone and weather lighter; at both base and top are layers of glauconite and quartz sand; 5 ft above base is a black bed, 3 ft thick.
	Abrupt contact		
Esopus		est 200	Siltstone, dk gray to black; noncalcareous; massive; well-developed slaty cleavage; breaks into small chips; generally lacking fossils.
Glenerie		20	Bedded chert, dark gray siliceous limestone, and shale; profusely fossiliferous, with silicified fossils locally present in brownish weathering residue.
Port Ewen		6 to 7	Limestones, argillaceous, med. gray; weathering brown; resembles New Scotland Formation.
Alsen		32	Limestone, fn to med gr, med bedded; with black chert in nodules and beds.
Becraft		46	Limestone, cs gr, med to thick beds; interbedded green shale seams in lower part; rare chert; weathers much whiter than adjacent units.

New Scotland Formation		98	Limestones, very argillaceous, shaly weathering, interbedded with dark gray calcareous shales; scattered chert nodules; contains large spiriferid brachiopod Maclopleura (formerly "Eospirifer") maclopleura and platyceratid gastropods.
Kalkberg Formation		54	Limestones, med gr, massive, cherty, with abundant crinoidal remains, bryozoans, and brachiopods; Dicoelosia (formerly "Bilobites") occurs beneath Thruway bridge about 12 ft below top; base drawn at lowest continuous chert layer.
Coeymans	Ravena	14	Limestones, med to cs gr, gray; in thick beds; chert-free.
Manlius	Thacher	9	Limestones, fn to med gr, dark blue.
		10	Stromatoporoid biostrome
		10	Limestones, fn gr.
		7	Stromatoporoid biostrome
		15	Limestones, dark blue, thin-bedded; some units laminated.
Rondout	Whiteport	4	Dolostones, fn gr, silty, thin bedded, weathering buff; some shale interbeds.
	Surface of unconformity		
Normanskill	Austin Glen	265 +	Interbedded gray siltstones and graded graywacke beds ranging in thickness from a few cm to many meters; graptolites abundant in some graded beds.

exposed in the glen, because only a small part of the total thickness is exposed and the basal contact is not visible, by modern standards, the choice of this place for a type locality is not satisfactory.

Estimates of the thickness of the Austin Glen Member range from 500+ feet (Ruedemann, 1942, p. 107) to 1200+ (Johnsen and Schaffel, 1967, p. B17). The thickness may be much more than this but has never been satisfactorily determined. Only the upper 265 feet (present author's measurement by tape and compass) are exposed in Austin Glen.

The chief point of stratigraphic interest attached to the Normanskill Formation is the controversy over its age and its association with the carbonate rocks of Ordovician age, which in various parts of the Hudson-Champlain Valleys were deposited contemporaneously with the Normanskill. In the area around Austin Glen no carbonate rocks of Ordovician age are exposed. Consequently the local exposures do not present the problem of determining the relationship between the Cambro-Ordovician carbonate rocks and the essentially noncarbonate Cambro-Ordovician terrigenous strata, a problem that must be confronted in numerous other localities in the Hudson Valley. This, the so-called "Taconic problem," has been a cause célèbre in American geology for 125 years. The problem now appears to have been solved (Zen, 1961, 1963, 1967; Sanders, Platt, and Powers, 1961; Platt, 1962; Theokritoff, 1964; Bird, in press). But, one of the reasons why it defied solution for so long was the controversy over the age of the Normanskill graptolite fauna. A short digression on the byplay that occurred between Rudolf Ruedemann and E. O. Ulrich on the subject of the correlation of the Normanskill is valuable in that it explains why progress in understanding the geologic relationships in the Hudson Valley was blocked for more than half a century.

When Ruedemann (1901a) first described the Normanskill fauna, he suggested that its age was Late Trentonian (in older sense; not Late Trentonian of G. A. Cooper, 1956). Ruedemann's reasoning was based on fossiliferous limestone pebbles in the Rysedorph Hill Conglomerate (Ruedemann, 1901b); some of these pebbles contain early Trentonian fossils. Because the Rysedorph Hill Conglomerate is intercalated within the Normanskill Formation, Ruedemann (1901b) concluded that the age of the Normanskill is post-lower Trenton. This conclusion brought Ruedemann afoul of Ulrich, whose stratigraphic interpretations were based on what Ulrich thought were well-established relationships among the Ordovician strata of northeastern Tennessee.

One of the correlations between New York and Tennessee upon which agreement was general is that the Normanskill fauna of New York occurs as well in the Athens Shale in the Appalachian Valley of northeast Tennessee. According to Ulrich's interpretation of the stratigraphic relationships in northeast Tennessee, the Athens Shale underlies a limestone which Ulrich took to be the same age as the New York Lowville. The New York Lowville is of pre-Trenton age; any unit that was pre-Lowville, therefore, was not only pre-Trentonian but also must be of Chazyan age, the same age as the Chazy Limestone underlying the New York Lowville. By this circuitous line of argument, Ulrich issued the fiat that the age of the Normanskill could not possibly be Trentonian, but actually was Chazyan! Subsequent work has shown that the unit in Tennessee correlated with the Chazyan actually is of post-Chazyan age (Cooper and Cooper, 1946, p. 60-62).

In the face of this dogmatic assertion by Ulrich, Ruedemann abandoned his original assignment of the Normanskill to the late Trentonian, and quoted Ulrich to the effect that the Normanskill is Chazyan. It is a curious twist of fate that so much of Ruedemann's career, especially his mapping, thus proved to have been a living lie. We can only wonder what private thoughts must have gone through Ruedemann's head each time he was forced to "knuckle under" in print to Ulrich by reciting that the age of the Normanskill is Chazyan. Perhaps Ruedemann soothed his own conscience by his persistent declaimers that the Normanskill might be as young as "Black River." How much farther along we would be if Ruedemann had been a more combative soul who could have stood up on his hind legs on this question and could have told E. O. Ulrich to go right straight to hell and could have made it stick!

As a result of a later sorting out of the graptolite faunas (Berry, 1960, 1962, 1963), it is now clear that Ruedemann's original age assignment of the Normanskill to the late Trentonian is correct and that in many localities the Normanskill Formation unconformably overlies carbonate rocks of early Trentonian and older ages. As long as the Normanskill was assigned to the Chazyan, however, it was necessary to infer a thrust at any contact where the Normanskill Formation occurred above carbonate rocks of post-Chazyan age. Because Ruedemann almost invariably showed a thrust at the western boudary of the Normanskill Formation, his mapping now becomes suspect and these contacts all require restudy.

<u>Rondout Formation, Whiteport Member</u>. Between Kingston and Catskill various units of sandstone, dolostone, and limestone of Late Silurian age and never reaching more than a few meters in thickness, occur between the base of the Thacher Member of the Manlius Formation and the surface of unconformity cutting the Austin Glen beds. In Austin Glen the only strata present above the Austin Glen and beneath the Thacher are about 1.5m of prominently cleaved, silty, buff-weathering dolostone. Following Rickard (1962) this silty dolostone is assigned to the Whiteport Member of the Rondout Formation.

<u>Manlius Formation, Thacher Member</u>. The distinctive "ribboned" limestones, stromatoporoid biostromes, and minor fine-grained dolostones that overlie the Rondout beds were previously assigned simply to the Manlius Limestone and considered to be of late Silurian age. As a result of Rickard's (1962) study, these carbonate rocks are now assigned to the Thacher Member of the Manlius Formation and are considered to be of early Devonian age. The base of the Devonian System, therefore, has been shifted downward from the top of the strata formerly designated as "Manlius" to the base of these strata. Why this boundary should be at the base of a particular formation and not somewhere within it has never been squarely faced; instead the

convenience is continued of supposing that any "well-behaved" systemic boundary "naturally" will occur at a plane where the lithologic characteristics change. In Austin Glen the Thacher is 51 feet thick (Rickard, 1962, p. 133).

Coeymans Formation, Ravena Member. In contrast with the Thacher strata, the overlying 14 feet of beds consist of medium-to coarse-grained, nonlaminated, gray calcarenites that are characterized by numerous individuals of the pentamerid brachiopod Gypidula coeymanensis Schuchert. These gray calcarenites, formerly simply the Coeymans, are now assigned to the Ravena Member of the Coeymans Formation (Rickard, 1962). As is explained under the section on sedimentology, the Thacher beds form the basal unit of a tripartite cyclic succession, of which two are present in the Lower Devonian beds. The cycles, from base upward are: (1) Ravena-Kalkberg-New Scotland; and (2) Becraft-Alsen-Port Ewen.

Kalkberg Formation. The Kalkberg Formation, 54 feet thick in its type locality, Austin Glen (according to the revised boundary of Rickard, 1962), is recognized by its fossils, chert, and silt content. The base of the formation is drawn at the lowest continuous chert layer. For the purposes of detailed mapping and geologic analysis of numerous borings, Dunn and Rickard (1961) found it convenient to adopt a twofold member subdivision of the Kalkberg Formation with each member having lower and upper divisions. Accordingly they proposed the Hannacroix Member, below (containing chert and about 25 per cent SiO_2 in chemical analyses), and the Bronks Lake Member, above (lacking chert and containing more than 50 per cent SiO_2 and several per cent of Al_2O_3 in analyses). The characteristic brachiopod of the Kalkberg interval is the dalmanellid genus Dicoelosia (formerly Bilobites).

New Scotland Formation. The top of the lower cycle of carbonate rocks is formed by the shaly weathering calcareous siltstones characterized by numerous brachiopods, of which Macropleura macropleura (Conrad) is most conspicuous. Small amounts of chert are present in the New Scotland Formation, but by comparison with the underlying Kalkberg, the New Scotland is not prominently cherty. The New Scotland is here used in the sense of Rickard, who has slightly altered the usage of Chadwick (1908) and earlier workers. The section of nearly vertical strata in the bed of Catskill Creek in Austin Glen is 98 feet thick (Rickard, 1962, p. 87; 132). This section looks to be undisturbed, but the presence of numerous bedding thrusts in the New Scotland Formation a short distance to the northeast suggests that the undisturbed appearance may be deceptive. The New Scotland beds are profusely fossiliferous but in Austin Glen specimen-collecting is nearly impossible (and probably also illegal, without a permit, according to New York State law). The original calcite of the fossils has been dissolved away, leaving exquisite impressions of the hard parts. The rock breaks so irregularly that it is not generally possible to remove these impressions. The recommended approach is to bring plasticene or liquid latex to the exposures and to collect replicas of the original skeletal materials. Presumably if the natural impressions are not damaged no permit is required for making such replicas.

Becraft Limestone. The base of the second cycle of carbonate rocks begins with a very coarse-grained, gray to pinkish, massive calcarenite characterized by abundant crescent-shaped (in section) remains of the crinoid Aspidocrinus scutelliformis Hall. This is the Becraft Limestone. The distinctive crinoid fragments are particularly abundant near the base of the Becraft, where the limestone beds are separated by greenish shale partings. Higher up the shale partings are not present, a little chert occurs, and the pentamerid brachiopod Gypidula pseudogaleata (Hall) becomes abundant. Rickard (1962, p. 89; 132) reported 48 feet of Becraft in Austin Glen, but near the natural dam, where the strata dip 84°, I have measured 63 feet. With bedding thrusts a distinct possibility, neither of these figures may represent the true thickness.

Alsen Formation. Gradationally overlying the Becraft is a unit, the Alsen Formation, which mimics the Kalkberg. The Alsen is finer grained than the Becraft and contains chert and much silt. In isolated exposures where the succession is uncertain, fossils are the only sure means of distinguishing the Alsen from the Kalkberg. The brachiopod Nucleospira concinna (Hall) and the bryozoan Monotrypa tabulata occur in the Alsen but are unknown in the Kalkberg (Rickard, 1962, p. 90). In Austin Glen Rickard (1962, p. 132) reported 32 feet of Alsen; my own measurement suggests 55 feet.

Port Ewen Formation. The counterpart of the New Scotland in the upper cyclic sequence is the Port Ewen Formation, which Rickard (1962, p. 132) found to be 6 to 7 feet thick in Austin Glen.

Glenerie Formation. Just at the point in Catskill Creek where the exposures become discontinuous is the place in the succession where the Glenerie Formation should occur. The Glenerie contains interbedded layers of chert, shale, and siliceous carbonate rocks; its thickness is about 20 feet in a roadcut exposure southwest of Catskill on N. Y. 23A just east of the Thruway, where the cuts give the impression that construction on an exit road was commenced but tha an exit here was later cancelled, so that the project was abandoned.

Esopus Formation. After encountering all the formations whose thicknesses are measured in a few tens of feet it is something of a drastic change to come upon the Esopus Formation, which consists of 200 to 250 feet of massive, much-cleaved, dark gray to black, noncalcareous siltstone. Brachiopods are sparse in the Esopus, thus probably explaining in part why it has not been subdivided in the past, But hope should not be abandoned; Boucot (1959) has described a distinctive Esopus brachiopod fauna from Highland Mills, N. Y. The Esopus forms the steep bluff on the northwest side of Catskill Creek, where the creek flows along strike between the southeast end of Leeds Gorge and the northwest end of Austin Glen.

Schoharie Formation. The Schoharie Formation is a transitional unit between the noncalcareous Esopus below and the Onondaga Limestone above. On the basis of the proportion of mudstone and interbedded limestones it is possible to recognize three members, from base upward Carlisle Center, Aquetuck, and Saugerties (Johnsen, 1959; Johnsen and Southard, 1962). Yellowish-brown weathering is a fairly distinctive feature of much of the Schoharie Formation. Other materials present are quartz sand, glauconite, and black, siliceous limestone that weathers black to reddish purple. In the Leeds Gorge section the Schoharie Formation is 69 feet thick (Table 1).

Onondaga Formation. The conspicuously cherty limestones containing abundant corals and other fossils that overlie the Schoharie Formation belong to the Onondaga Formation. Detailed studies permit the recognition of various members of the Onondaga, three of which (from base upward, the Edgecliff, Nedrow, and Moorehouse) occur in the Leeds Gorge section (Oliver, 1956, 1962). About 115 feet of Onondaga are present in this section but the topmost layers are not present. The overlying Bakoven Shale, Mount Marion Formation, and other higher Devonian formations will not be seen on this trip.

Sedimentology

Of the many aspects of the sedimentology of the two contrasting suites of strata to be seen in the "Little Mountains," the one to be most emphasized here is the presumed effect of water depth. The characteristics of the Middle Ordovician Austin Glen strata are thought to have resulted from deposition in deep water (depths in thousand of meters) with depth changes and migrations of the shoreline during sedimentation exerting negligible effects on the sediments. By contrast the characteristics of many of the Silurian and Devonian strata are thought to have originated as a result of deposition at the margin of the sea, in shallow water (a few tens of meters) close to sea level, with depth changes and migrations of the shoreline during sedimentation exerting major controls on the distribution of materials. For those interested in understanding the kinds of materials implied by the term flysch, reference is made to characteristics of the Austin Glen Member of the Normanskill Formation.

Characteristics of Austin Glen strata

The Austin Glen beds to be examined at Stop 1 (Fig. 1) consist of nearly equal parts of siltstone and graywackes. Not only is the proportion about equal, but the distribution of bed thicknesses is comparable. The thickness of siltstone layers ranges from a fraction of a centimeter up to the class limited by 64 and 128 cm; the median falls within the 8 to 16cm class. The graywackes range up to the class limited by 128 and 256 centimeters, with a median in the 8 to 16cm class.

Sedimentary structures are numerous and can be employed to ascertain the original top direction of the strata, the direction of flow of the

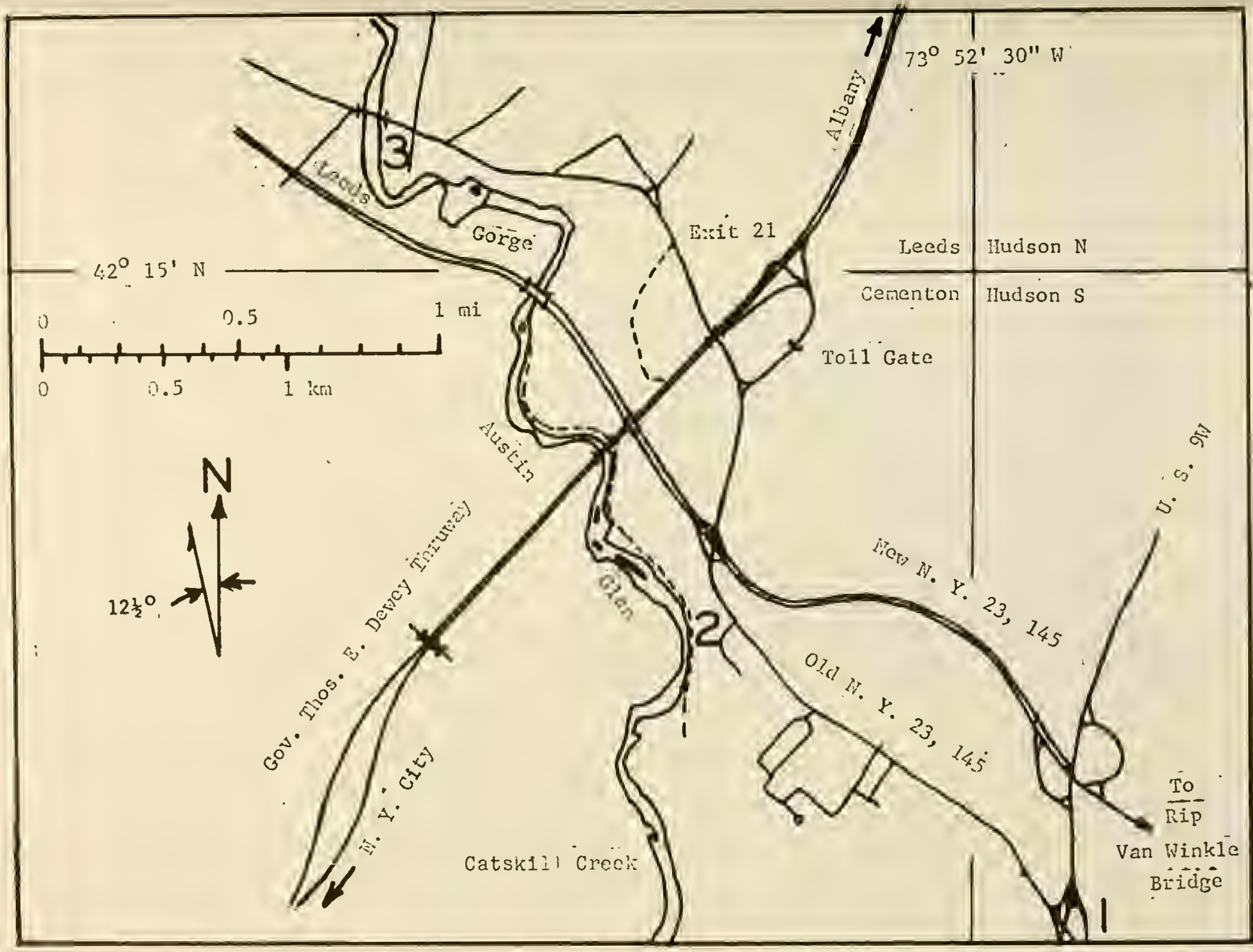

Fig. 1. Location map for Austin Glen and Leeds Gorge. Names at intersection of 42°15' N and 73°52'30" W are of U.S. Geol. Survey 7 1/2-minute topographic quadrangle maps. Numbers show trip localities: 1, in Austin Glen beds on old N. Y. 23-145, Catskill; 2, start of trip in Austin Glen; and 3, end of trip in Leeds.

depositing currents, and the interactions between the moving current and sediment being deposited.

Geopetal criteria on display in the Austin Glen beds include grading, truncated cross-laminae (not many of the cross-laminae present can be used to determine tops; numerous examples are present in which sediment fallout accompanied current flow with the result that the ripples built upward as they migrated downcurrent and the cross-laminae were not truncated), sole markings (not very abundant in the sections to be seen on this trip), and a feature not previously dignified by having appeared in the literature but which I have found useful and have dubbed "glazed bottoms." Glazed bottoms accompany normal upward grading and seem to have resulted from the acquisition, by the coarse particles in the base of the bed, of a veneer of fine particles from below. The light reflecting from the fine particles creates a sheen resembling glazed pottery. Glazed bottoms accompany the well-developed parting that occurs at the base of a graded bed, in contrast with the indistinct parting at the top where gradation exists into the overlying siltstone. Even where distinct partings occur at both top and bottom, I have found "glazing" only on the bottom surface, hence the name.

The paleocurrents responsible for the Austin Glen graywackes near Catskill flowed from chiefly W to E and from NW to SE; a few flowed from SE to NW. How these directions can be rationalized with the regional paleogeography of Mid-Ordovician time in the Hudson Valley is not yet understood. Reconnaissance paleocurrent studies have been made both to the north (Middleton, 1965) and to the south (McBride, 1962), but these measurements lack a detailed stratigraphic base. Presumably off to the west most of North America was submerged by the Trenton sea in which carbonate sediment was accumulating. The regional submarine slopes were to the east. Yet during Normanskill time, a monumental slope reversal must have occurred for into the Normanskill sea from the east came a-sliding the vast body of Taconic strata, which originally accumulated yet farther eastward (see summary in Bird, in press).

Paleocurrent indicators in Austin Glen itself include an extraordinarily large mold of a flute (60cm long, 30cm wide, and 8cm deep) exposed on one of the thick beds by the old millrace, and large-scale cross-strata (Fig. 2). These large-scale cross-strata, first noticed by Schuchert and Longwell (1932, p. 316), are fine examples of the concave-up variety that is truncated at the top and tangential at the base. They are distinctly anomalous among these graded beds. They occur at the top of a bed graded that is 4m thick and fill a depression 0.5m deep that is exposed for 2.7m in the direction parallel to current flow (here straight up the dip, or from NW to SE). At extreme low water an even larger depression cut into the top of this bed is exposed. This larger depression is 15m wide, 0.7m deep; in it the dip of the cross-strata parallels the sides of the depression.

At stop 1 are exposed some multiple wisp-like features ("flame structures" of authors) that clearly demonstrate that the current was not

Fig. 2. Large-scale cross-strata filling shallow channel cut in top of a thick graded bed in Austin Glen Member of the Normanskill Formation, northeast side of Catskill Creek, Austin Glen, on northwest limb of tight anticline, at old millrace. Direction of paleocurrent flow was directly up the dip, or from NW to SE. Such large-scale cross-strata are rare in sequences of graded beds such as the Austin Glen Member. The origin of the currents responsible for scouring the channel and for filling it with coarse, cross-stratified sand is not known. (Photo by J. E. Sanders, April, 1959.)

only bringing along the graded sediment but simultaneously was dragging up wisplike structures composed of the underlying material. The wisps were supported by the sediment being deposited around them.

Other features of interest in the Austin Glen Member include two sandstone dikes, exposed along Catskill Creek bank several hundred meters downstream of the point where the path down from Austin Acres intersects the abandoned railway grade.

No attempt is here made to cite the voluminous literature that exists on the kinds of strata exemplified by the Austin Glen Member. A few of the published studies of the Austin Glen rocks however, have been included (Middleton, 1962, 1965; Weber and Middleton, 1961).

Although a few dissenters may still exist, the preponderance of geologic thinking today is that the modern counterparts of the Austin Glen Member are to be found in the abyssal plains that carpet the floors of the oceans. More particularly the abyssal plains of today that would exemplify the conditions that can be reconstructed as having prevailed during Ordovician time are located within a few of the great marginal trenches.

Characteristics of Silurian-Devonian strata

In contrast with the noncalcareous, graptolite-bearing Austin Glen terrigenous strata, the Silurian and Devonian Formations typically are calcareous and contain faunas consisting chiefly of brachiopods, bryozoa, and corals. Facies studies, principally by Oliver, Rickard, and Laporte, have emphasized the concept that the variations encountered in a vertical stratigraphic succession are a reflection of the lateral distribution of sedimentary environments which prevailed at time of deposition. This important re-affirmation of the principle known as Walther's Law is based on careful analyses of both the faunas and the physical characteristics of the strata. The variations occur both on a large scale among formational units and on a small scale within formations.

Large-scale variations. The cyclic arrangement of the two groups of three Helderbergian formations (Coeymans, Kalkberg, New Scotland, and Becraft, Alsen, Port Ewen) has been noticed by numerous observers. Rickard (1962) adopted the explanation that this cyclicity resulted from variations in water depth during the Helderbergian submergence. He suggested that in the Helderbergian sea three nearshore facies belts had been present. From the shore outward he named these Neritic A, Neritic B, and Neritic C. In the A belt coarse-grained carbonate sands accumulated. These passed seaward into muddy and silty carbonate sands of belt B. Finally, in belt C the sea floor was composed of silts and muds. As submergence and progradation occurred, these belts shifted back and forth across the Hudson Valley, leaving behind different sediments. A gradual deepening would result in successive deposition, at a single point, of facies A then facies B and finally facies C. These would overlie one another in a fixed sequence, from bottom upward A, B, C. Two such cycles of gradual deepening appear to be represented by the Coeymans (A), Kalkberg (B), New Scotland (C) and Becraft (A), Alsen (B), and Port Ewen (C) cyclic successions. If this interpretation is correct, as seems probable, then we must infer that a marked shoaling occurred at the contact of the New Scotland with the Becraft.

Applying this same line of thought to the Glenerie-Esopus-Schoharie-Onondaga succession, we conclude that the deepening implied by the Port Ewen strata continued, and probably reached its maximum at the time the Esopus was deposited. The Schoharie evidently resulted from a gradual shoaling; its bathymetric affinities are with the New Scotland and Port Ewen. The Onondaga evidently records a return to shoal-water conditions, but with an important change from the conditions of the Coeymans and Becraft: coral bioherms were abundant.

It may seem paradoxical that the strata deposited in supposedly deepest water contain the most numerous stratigraphic hiatuses. I see nothing unusual in this relationship provided one will admit the possibility that for various reasons and for various lengths of time on the sea floor, little or no sediment may accumulate in a given area. Most of the discontinuities in the succession that have been described are characterized by zones containing glauconite or phosphatic material (for example, Goldring and Flower, 1942). Accordingly, I infer that these gaps resulted from submarine conditions and did not involve subaerial exposure, a condition commonly implied by the use of the term "unconformity" to describe the stratigraphic relationships.

Small-scale variations. The Helderbergian cyclic changes, just described, from coarse calcarenites to fine mudstones, must have involved variations in water depth of perhaps tens of meters. The sediments involved do not imply any unusual salinities nor proximity to coral bioherms. By contrast, some of the small-scale variations displayed in the Manlius and Onondaga formations suggest depth variations of only a few meters and do involve higher-than-normal salinity (at least of the interstitial waters) and the presence of coral bioherms. In his close study of the Manlius Formation Laporte (1964a, 1964b, 1967; also Rickard, Oliver, and Laporte, 1963) has found indications that deposition was controlled by proximity to sea level under climatic conditions that were warm and dry enough to cause evaporation of sea-marginal interstitial waters and thus to elevate the salinities of these waters to the point where dolomite would precipitate. By analogy with modern carbonate depositional provinces Laporte has recognized three sea-marginal facies: (1) supratidal, (2) intertidal, and (3) subtidal. In the supratidal environments were deposited laminated dolomitic mudstones that commonly display mudcracks and "birdseyes" and that contain rare fossils. These include algal mats, ostracods, and burrow structures. Examples occur in the middle and upper Thacher Member.

Intertidal environments were sites wherein collected pellets, large particles of lime mud; the resulting products were "ribboned" limestones and limestone-pebble conglomerates. A few kinds of fossils occur in these strata, each kind being represented by enormous numbers of individuals. Examples are ostracods, tentaculites, brachiopods, algal stromatolites, and oncolites. Products of intertidal environments occur in the lower Thacher Member.

In the subtidal environments were found pellets and carbonate skeletal remains, including biostromes. Abundant fossils include stromatoporoids, corals, codiacean algae, brachiopods, ostracods, and gastropods. Examples occur in the middle and upper Thacher. Laporte (1967) inferred that these sea-marginal sediments of the Manlius Formation accumulated in a lagoon that was protected from the open sea by a barrier consisting of carbonate skeletal material that eventually formed the Coeymans Limestone.

Oliver's (1956, 1960, 1963; also Rickard, Oliver, and Laporte, 1963) study of the Onondaga Formation reached the conclusion that the several members had been deposited under varying conditions of water depth and proximity to coral bioherms. The proportions of various kinds of skeletal material varied systematically with environment; as the environmental boundaries shifted the kinds of sediment accumulated changed. Accordingly the subdivisions of the Onondaga Formation are based on variations in the proportions of skeletal debris, chiefly from corals, brachiopods, and echinoderms. An additional variation has been introduced subsequently; chert has formed.

In addition to these sedimentologic and ecologic analyses, the Silurian and Devonian strata have been scrutinized chemically and mineralogically. Published reports include Dunn and Rickard (1961), Fenner and Hagner (1967), Fessenden (1960), and Lindholm, (1969); in addition there have been at least 4 doctoral dissertations completed that have not yet been published (see list at end of references).

GEOLOGIC STRUCTURE

The structural features of the area include folds, various kinds of faults, regional joint systems unrelated to local structural features, calcite-filled fractures related to local structural features, and cleavage. The ensuing discussion of regional joints and cleavage is based largely on the unpublished detailed study by Babcock (1966ms). Following the description of the structures is a discussion of the age of the deformation. The major structural features are shown on a tectonic map (Fig. 3).

Folds

Despite its narrow width of 2km near Catskill Creek, the "Little Mountains" fold belt displays two contrasting longitudinal parts that are separated by the persistent Leeds anticline (Babcock, 1966ms), whose northwest limb locally has been broken by a thrust fault (Fig. 4). The eastern part contains generally older formations (typically Becraft Limestaone and older units); in it the anticlines are narrow whereas the synclines are broad and flat-bottomed. This configuration strongly suggests that the folds are related to places where bedding thrusts cut diagonally across the strata from one slip surface to another (Rich, 1934).

By contrast, the western part of the fold belt contains younger strata (a common surface unit is the Onondaga Limestone), and its folds included broad, open anticlines and narrow, tight synclines.

Whether this subdivision, apparent near Catskill Creek, persists throughout the "Little Mountains" remains to be determined. At any rate, in the trip area it is a reality and as a result, the Ordovician strata lie much closer to the surface in the eastern part of the fold tract. Typical projected depths to the Ordovician in the eastern part (ignoring possible bedding thrusts) lie in the range of 30 to 150m whereas in the western part typical projections range from 210 to 300m.

The cross sections of Fig. 4 do not attempt to portray the bedding thrusts. The scale of these sections (1:24,000) is much too small for this purpose. In order to display the bedding thrusts it is likely that both maps and sections would have to be on a scale of the order of 1:2400, or 10 times that of Fig. 4.

Faults

The faults include thrusts, normal faults, reverse faults, and strike-slip faults.

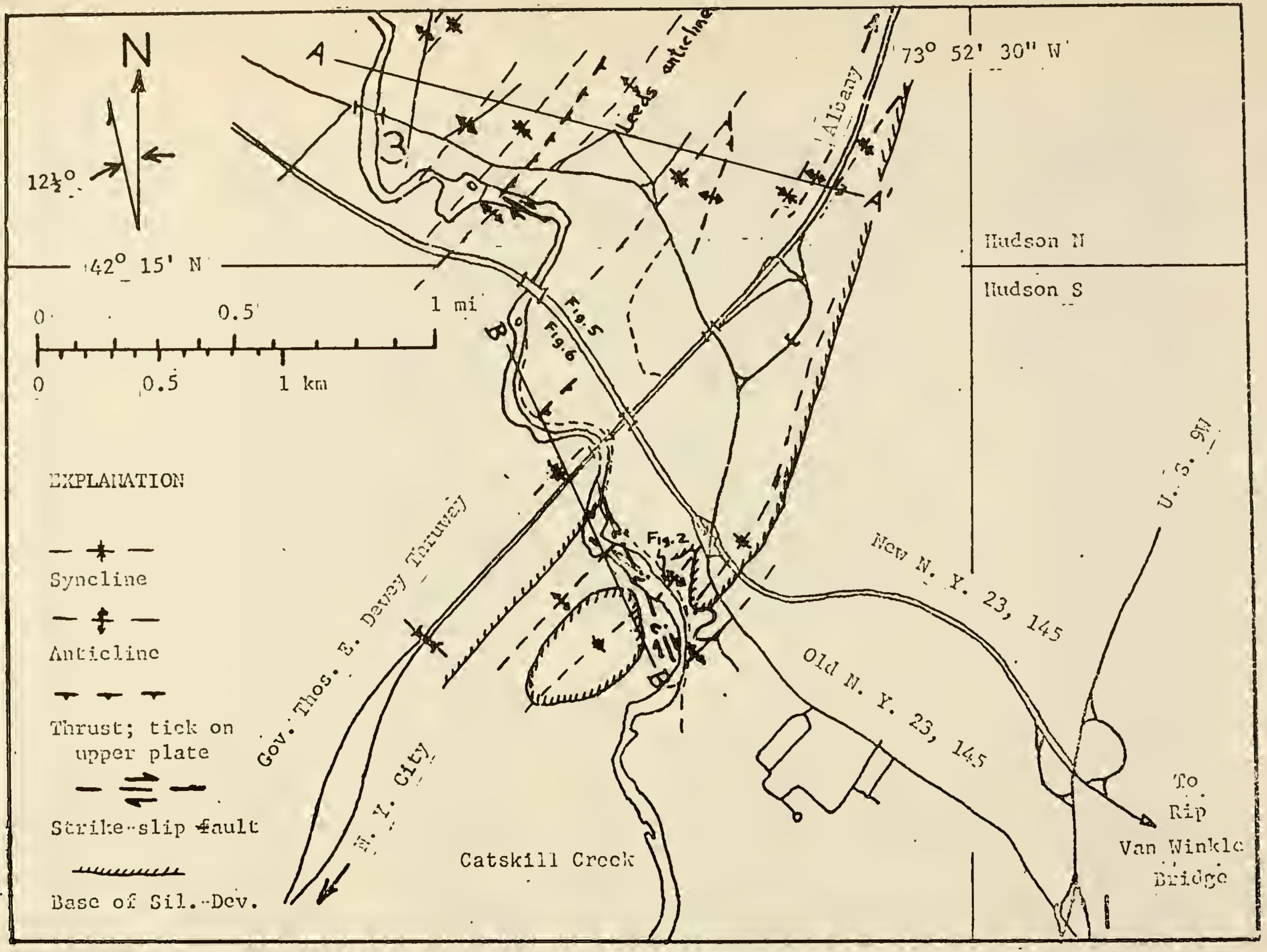

Fig. 3. Tectonic map of "Little Mountains" adjacent to Catskill Creek. Mapping on Leeds 7 1/2-minute quadrangle by E. A. Babcock 5th (1966ms); mapping on Cementon 7 1/2-minute quadrangle by G. H. Chadwick (1944), revised by J. E. Sanders, April, 1969. Map also shows locations of photographs (Figs. 2, 5, and 6) and cross sections (Fig. 4).

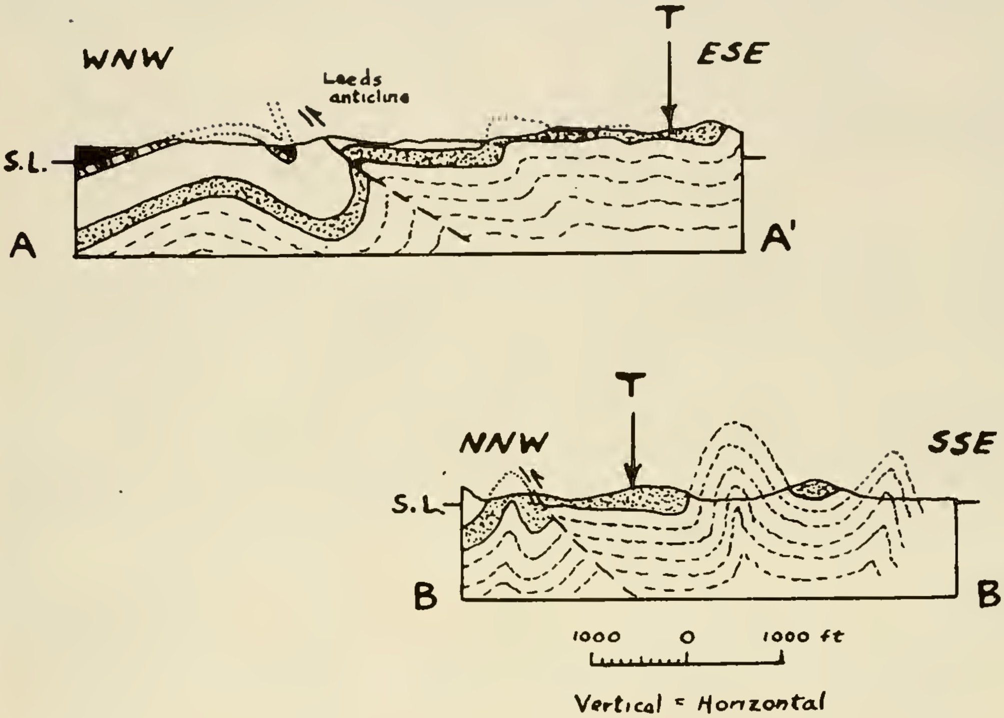

Fig. 4. Profiles and sections through "Little Mountains" near Catskill Creek, northwest of Catskill, N. Y. AA' on Leeds 7 1/2-minute quadrangle (after Babcock, 1966ms). BB' on Cementon 7 1/2-minute quadrangle. Capital T marks location of Thruway. Dashed lines, Austin Glen Member; Stippled, Rondout through New Scotland; white, Becraft through Schoharie; rectangles with black dots, Onondaga Formation; black, Bakoven Shale. Bedding thrusts omitted. Fig. 3 shows locations of AA' and BB'.

Thrusts

Within the area are examples of thrusts associated with broken folds and of bedding thrusts. Possibly this distinction is an artificial one and is not genetically significant; but in considering the geometric relationships between displaced strata and the fault surfaces, it is a useful subdivision to make.

Thrusts associated with broken folds. Thrusts associated with broken folds repeat the stratigraphic succession by bringing together from adjacent folds, limbs or axial parts that were not originally contiguous. The duplications of the strata occur at the expense of cutting out parts or all of the broken limbs of the fold(s). Geometrically, thrusts associated with broken folds and the strata they displace show various configurations, depending on where the limb of the fold broke and on the amount of displacement. On both upper and lower fault blocks the thrust may parallel the strata or cut across the strata at various angles. The thrust surface may become folded or even overturned, either before or after the overturning of any associated overturned strata. Overturned strata from the limb of an overturned fold can be thrust against right-side-up strata of the normal limb.

Ordinarily, thrusts associated with broken folds are easily recognized because the thrusts displace features that result from folding, such as cleavage and bent strata. Also, thrusts may pass laterally into nonbroken fold limbs. In some cases, however, the ultimate arrangement of thrusts associated with broken folds conceals the folds. An example is the series of monoclinal faulted slices of repeated, right-side-up belts of strata found in the Appalachian Valley in northeast Tennessee, so vividly described by Bailey Willis (1893, p. 226-228).

Nearly all previous investigators have assigned all the thrusts of the "Little Mountains" to the category of thrusts associated with broken folds (van Ingen and Clark, 1903; Dunn and Rickard, 1961; Babcock, 1966ms, for example). Clearly many of the thrusts in the "Little Mountains" are indeed of this kind; some pass laterally into nonbroken folds (Babcock, 1966ms) and others display the various characteristic geometries showing how the thrust broke across some part of a fold (Chadwick, 1910).

Bedding thrusts. By contrast with thrusts associated with broken folds, bedding thrusts display rather limited geometric relationships with displaced strata. Characteristically bedding thrusts initially cut across strata before the strata become folded, hence without regard to folds. As a matter of fact, by their very displacement bedding thrusts give rise to distinctive folds (Rich, 1934). Of course, the limbs of such folds may break and give rise to still other thrusts (Gwinn, 1964). The characteristic feature of bedding thrusts, as brilliantly diagnosed by J. L. Rich (1934), is that the strata on the lower block, which may be everywhere flatlying, parallel the thrust except where the thrust surface cuts across the strata in changing from one slipping level parallel to bedding to another, stratigraphically higher, slipping level parallel to bedding. At such places of cutting across, both the thrust itself and the strata on the upper block are not parallel to the strata on the lower block. Instead the strata on the upper block are parallel to the thrust surface. On the upper block, however, at all points forward of the place where the thrust changes from a locus of cutting across to a position parallel to the bedding, hence above places where the thrust parallels the strata on the lower block, the strata are inclined <u>toward</u> the thrust surface and terminate against this surface.

A characteristic geometric relationship of bedding thrusts is that they duplicate only parts of the stratigraphic succession and elsewhere, because they involve only slippage along bedding surfaces and create little or no change in stratigraphic thickness, they tend to remain obscured. The extent of duplication serves as a measure of displacement on the thrust; strata are repeated only in those zones between the intersection of thrust and strata on the lower block and the termination of the strata of the upper block against the thrust. Although some bedding thrusts originate in horizontal strata and large parts of such thrust surfaces may remain horizontal even after displacement, other bedding thrusts become folded (Figs. 5,6) and overturned after displacement.

The "folds" created by displacement on a bedding thrust simply die out in the direction where the amount of displacement decreases.

Bedding thrusts may duplicate strata through long distances without displaying any visible geometric relationship of a thrust cutting across beds.

The common occurrence of bedding thrusts in the "Little Mountains" has not been recognized previously. Although the details remain to be worked out, I am convinced that many of the geometric relationships between strata and thrusts in the "Little Mountains," previously attributed to the effects of thrusts originating from broken folds, will prove to be products of the folding of bedding thrusts.

<u>Normal faults</u>

In the "Little Mountains" are steep-dipping normal faults, that generally parallel the structural trend and cut across all other structural features. The normal faults in the folded belt have not been much discussed, although Goldring (1943, p. 303) suggested that in the Coxsackie quadrangle most faults are "of the normal type." Chadwick (1917) proposed that a genetic connection exists between the thrusts in the Ordovician strata and the later normal faults, but this idea has not attracted much of a following.

Fig. 5. Three bedding thrusts that have been folded, northwest side of new N. Y. 23-145, between Catskill Creek and Gov. Thomas E. Dewey Thruway (location shown on Fig. 3). New Scotland Formation on upper three thrust plates has been thrust over the Alsen Formation (in axis of anticline). New Scotland beds overlying lowest thrust strike into horizontal segment of lowest thrust that is just out of the view in lower left-center. Lowest thrust in this view is inferred to be the highest thrust of Fig. 6. (Photographs by Daphne diSomma, April, 1969.)

Fig. 6. Two folded bedding thrusts (and subsidiary thrust) bring New Scotland Formation over the Alsen Formation. Southwest side of new N. Y. 23-145, opposite from Fig. 5 (see Fig. 3.). Highest thrust in this view is inferred to be the lowest thrust in figure 5. (Photographs by Daphne diSomma, April, 1969.)

Reverse faults

In Leeds Gorge are exposed a few reverse faults of small displacement (Babcock, 1966ms, fig. 8). Generally no distinction can be made between reverse faults and steep-dipping thrusts associated with the limbs of folds.

Strike-slip faults

Goldring (1943, p. 304) found small faults with two trends; NE-SW and NW-SE which she characterized as being cross-faults. Likewise Babcock (1966ms) mapped a small NW-SE-trending fault in Leeds Gorge, against which a thrust stops. Both of these are here considered to be small strike-slip faults. In describing these faults as of the strike-slip variety the writer does not wish to convey to the reader visions of grandeur on the scale of the San Andreas fault of California. Perhaps tear-fault would be the best term; however I prefer strike-slip because of its connotations with respect to the displacement of vertical reference surfaces.

I suggest that a small strike-slip fault trending N25W passes through Austin Glen. I have shown such a fault with dashed lines and question marks on the tectonic map (Fig. 3). This is based on what appears to be the offsetting of vertical-axis folds. The isoclinal, upright anticline in the Austin Glen beds at the old millrace extends across Catskill Creek without displacement, as indicated by the continuity of the riffle-making ledges. However, farther down the strike to the southwest rises the synclinal ridge bringing down the Lower Devonian carbonate rocks. I suggest that this synclinal ridge is the offset part of the synclinal structure that is passed in entering Austin Glen from Jefferson Heights (via Austin Acres and the old footpath by the spring). If so, then the continuation of the millrace isoclinal anticline would be found just northwest of this synclinal ridge; it would have to be the anticline exposed just southeast of the Thruway Bridge. The Thruway Bridge anticline clearly is continuous across Catskill Creek and beyond to the southwest but its extension northeastward is not suggested by the topography. If this anticline by the Thruway Bridge is the same structure as the anticline at the old millrace, then the small strike-slip fault that offsets the axis must cross Catskill Creek in the covered interval where no exposures are present. If the structure is as I suggest, then the strike-slip offset has been 0.3km in a right-lateral sense. The detailed geologic mapping to test the validity of this structural interpretation has not been done at the time of writing of this guidebook.

Regional joint systems unrelated to local structural features

Babcock (1966ms) showed that in the "Little Mountains" near Leeds 4 sets of regional vertical joints are present and that these joints not only are unrelated to the folds but also originated prior to folding. The average strikes of these 4 sets of joints are: N 12°E, N 82°W, N 37°E, and N 69°W. These sets of joints display practically the same orientation as the sets of regional joints measured on the Allegheny Plateau by Parker (1942). Just as in the "Little Mountains," the regional joints in the Allegheny Plateau are unrelated to and originated earlier than the folds.

Calcite-filled fractures related to local structural features

Many calcite-filled fractures occur near and along faults (Figs. 5 and 6). Many of these fractures are characteristic en echelon "gash" fractures that occur perpendicular to the direction of maximum tensile stress in zones where shearing movements were concentrated. These calcite-filled fractures show various orientations and all clearly originated during the folding and faulting of the strata.

Cleavage

Babcock's observations demonstrate that the pronounced cleavage, so conspicuous in the massive siltstones of the New Scotland, Esopus, and Schoharie formations, is not only a regional feature but also that the cleavage is closely connected with the folds. The regional cleavage parallels the strike of the fold axes and is inclined at the same average dip as that of the axial planes of the folds. North of the valley of Catskill Creek Babcock (1966ms) found that the average strike of the regional cleavage is N 18°E and its average dip is 82°E. In the valley of Catskill Creek the average strike of cleavage (and of fold trends) is N 40°E and the dip of the cleavage varies because on some fault blocks the cleavage has been rotated away from its average dip of 82°.

The regional cleavage affects not only the Silurian and Devonian formations, but also the Ordovician strata (Pepper, 1934). Within the "Little Mountains," at least, only one cleavage is present and its age is the age of the deformation of the Devonian and older strata.

In many exposures the cleavage is the most prominent parting. In fact, when first examining an exposure of siltstone in the "Little Mountains" one would do well to suspect that the most obvious parting is the cleavage.

In the south wall of Leeds Gorge, an anticline in the Esopus Formation is exposed in complete transverse section. At this place the relationship between axial-plane cleavage and the strata of the fold is well displayed.

Little or no metamorphism has occurred along cleavage surfaces near Catskill Creek. However Babcock (1966ms, p. 35) reports that 1500 feet east of Limestreet is an exposure of intensely deformed New Scotland Formation where a phyllite has been created.

Age of deformation

The "Little Mountains" were first celebrated not only as small-scale extensions from Pennsylvania into New York of the Appalachian Valley and Ridge Province, but also as examples of structures formed within the Appalachian orogeny late in the Paleozoic (W. M. Davis, 1882, 1883; van Ingen and Clark, 1903; Ruedemann, 1942). Because of the structural similarities between the "Little Mountains" and the full-scale Valley and Ridge Province, nearly all early workers favored a late Paleozoic date for the final deformation in the Hudson Valley. However, the local stratigraphic evidence on this date is ambiguous and not subject to a unique interpretation. No new data have yet changed the long-standing fact that the youngest deformed strata in the "Little Mountains" are of medial Devonian age. Therefore (as emphasized by Schuchert and Longwell, 1932, p. 323; Chadwick and Kay, 1933, p. 7; Rodgers, 1967b, p. 416; and others) the final deformation in the Hudson Valley could have occurred any time after the middle of the Devonian Period.

Earlier workers tended to emphasize the majesty and grandeur of the effects of the Appalachian orogeny of Late Paleozoic age, which they took as a monumental revolution that brought down the curtain on the Paleozoic Era. Although he was instrumental in fostering this concept in beginning students, Schuchert (1930) started to reverse this trend in the Hudson Valley by calling attention to the importance of the Acadian orogeny of Devonian age. Of late the pendulum thus set in motion has swung the other way. The Appalachian orogeny has come upon evil days. It has become fashionable to ridicule the significance of this supposedly majestic orogeny. Because of their small size and nearly north-south trend, and because their youngest deformed strata are of medial Devonian age, as noted above, Woodward (1957a, 1957b) has stripped the "Little Mountains" away from the Appalachian orogeny and has reassigned them to the Acadian. In his attempt to "think Acadian and help stamp out Appalachian" Woodward has meted out the crowning humiliation; he has suggested that the name of the late Paleozoic deformation in the Appalachians be changed from Appalachian to "Alleghanian orogeny."

I think that the evidence from the "Little Mountains" does not support Woodward's line of argument; rather I think the evidence reaffirms the essential correctness of the W. M. Davis interpretation. The stratigraphic evidence is as ambiguous as ever, hence the case in defense of Davis, just as the case made by Woodward, rests on structural considerations that are not conclusive. These structural considerations are: (1) the small size of the folds in the "Little Mountains" as compared with the large size of the folds in the Appalachian Valley and Ridge Province in Pennsylvania; (2) the nearly north-south trend of the fold axes in the "Little Mountains" as contrasted with the nearly east-west trend of the Appalachian folds in northeastern Pennsylvania; (3) the "style" of deformation in the "Little Mountains" compared with that of the Appalachian Valley and Ridge Province in Pennsylvania; and (4) the similarity in orientation and in relationship to folds of regional joint systems in both areas.

Size of folds

The difference in size between the folds of the "Little Mountains" and those of the Valley and Ridge Province has been variously interpreted. Shaler (1877), Davis (1882, 1883), and others supposed that the "Little Mountains" were simply miniature northward extensions of the larger Valley and Ridge folds. By contrast, Chadwick (in Goldring, 1943, p. 288) and Woodward (1957a; 1957b) have argued that the difference in size means difference in age.

Just what is this difference in size and what is its significance?

The largest folds in the "Little Mountains" are at least an order of magnitude smaller than the largest folds in the Appalachian Valley and Ridge Province in Pennsylvania. A typical amplitude for folds in the "Little Mountains" is 0.4 to 0.75 km (Babcock, 1966ms), whereas in Pennsylvania it is 8 to 16 km (Gwinn, 1964).

The "little" folds are not distinctly set off from the "big" folds. On the contrary, southwest of Kingston, New York the "little" folds become "big" folds. From the Shawangunk Mountains southwestward, the amplitudes of the folds become progressively larger. I regard it as more than coincidental that this change in size of folds occurs exactly at the place where the Shawangunk Formation, a tough, competent sandstone and conglomerate, pinches out against the surface of unconformity that truncates the Ordovician strata. Not only the Shawangunk, but also other Silurian formations (Schuchert, 1916; Chadwick and Kay, 1933, p. 3, fig. 3) the high Devonian strata, and even some of the lower Devonian strata, thicken notably southwest along the strike of the present outcrop belt. Where thick, competent units are present in the succession, the folds are "big"; by contrast where thick, competent units are not present in the succession, the folds are "little". I think the matter of size of folds resolves itself to this stratigraphic factor. If so, then the contrast in size of folds between "Little Mountains" and central Pennsylvania is not an argument that supports different times of deformation. Rather, insofar as age of deformation is concerned, size of folds becomes irrelevant and size indicates only a contrast in thickness and competency of units deformed.

Trend of folds

Tectonic analysts depend to a great extent on the principle that individual orogenies tend to create characteristic trends of folds. In a given region showing two discrete trends of folds, therefore, one reasonably suspects that two orogenies have occurred. In suggesting that the "Little Mountains" were deformed in the Acadian orogeny, Woodward (1957a, 1957b) argued that the prevailing nearly north-south trend of the "Little Mountains" strikes nearly at right angles to the trend of the Appalachian folds, which Woodward inferred strike east-west across southern New York. On the basis of these supposedly discrete trends, Woodward argued that two orogenies had occurred. Because he regarded the east-west trend as "Appalachian" (his "Alleghanian"), he supposed that the north-south trend must be Acadian. Some of the change of trend takes place at Kingston, (Rodgers, 1967a), where the folds change size.

If two such separate trends do exist, as supposed by Woodward, they might serve as a strong argument in favor of two orogenies. But it must be added, that the trends themselves still do not prove whether the orogenies were of Acadian, Appalachian, or Jurassic age. As far as I am concerned, the burden of proof still rests with those who insist that two discrete structural trends exist. In the Hudson Valley I have not seen any evidence that could not also reasonably be interpreted as merely culminations and depressions with axes transverse to a single main fold trend.

But suppose the east-west trend cited by Woodward is real and indicates a second orogeny that is younger than the north-south trend? What is the age of the folds with the east-west trend? In eastern Pennsylvania the folds involve the Paleozoic strata (whatever age their youngest member is), but the change of trend involves in addition the Upper Triassic Newark strata. The age of the Paleozoic strata folded in eastern Pennsylvania sheds no more precise light on the age of the east-west trend there than the medial Devonian age of the youngest strata deformed in the "Little Mountains" sheds on the age of the deformation of the north-south trend in the Hudson Valley. I intend to argue the case more fully elsewhere, but nevertheless it is pertinent to add here that in eastern Pennsylvania, the Triassic strata, just as the Paleozoic strata, change trend from northeast-southwest to nearly east-west. Furthermore, where the Triassic strata change strike, their dips become nearly vertical, whereas these dips rarely exceed 25° elsewhere where the strike is more nearly northeast-southwest. I construe this evidence to mean that after deposition of the Newark strata, the entire Appalachian belt was bent, probably in the Jurassic (deBoer, 1968), as an orocline in the sense of Carey (1953, 1958). This zone of bending follows lat 40°N, a major lineament whose importance has been emphasized by Woodward and Drake (1963). The similarity of trend between Triassic and Paleozoic strata suggests an interpretation not discussed by Woodward and Drake. The lack of difference in trends between Paleozoic and Triassic strata suggests that on the continental block the major deformation occurred after Late Triassic time and prior to Late Cretaceous time. The greater offsets of features on the ocean floor as compared with those on the continental block may be the expression of a long-active transform fault (Wilson, 1965), which experienced major displacements during Cenozoic time.

If the age of bending of the Appalachians is indeed Jurassic as suggested above, then the east-west trend of the folds in eastern Pennsylvania does not prove an Acadian age for the north-south trend of the "Little Mountains" of the Hudson Valley. If two fold trends are really present in the sense of Woodward and if these trends indicate two orogenies, then these fold trends could as well be products of Appalachian and Jurassic orogenies as of Acadian and Appalachian orogenies. In short, the argument based on the supposed two fold trends, even if these two trends really do exist, is not conclusive as Woodward supposed.

Style of deformation

The question of "style" of deformation is not an easy one to resolve. It becomes all the more difficult when one is dealing with superimposed effects of several orogenies. Furthermore, although we may agree on the "style" of the Appalachian orogeny, how are we to decide what was the "style" of the Acadian orogeny? I do not know what may have been the "style" of the Acadian orogeny in the Hudson Valley. I do contend that the "style" of the Appalachian orogeny is distinctive and furthermore that new evidence from the Hudson Valley supports the concept that the "style" of deformation in the "Little Mountains" exactly matches that of the diagnostic Appalachian "style". In this coincidence I see a substantial reason for considering the age of the deformation of the "Little Mountains" to be Appalachian rather than Acadian. Two aspects of this "style" to be discussed are: (a) the presence of bedding-plane thrusts, and (b) the way in which the folds die out northward and westward.

(a) Bedding-plane thrusts. Nearly all tectonic philosophers meditating about the Appalachians have concluded that a "guide style" to the Appalachian Valley and Ridge structure, hence to the Appalachian orogeny, is what has come to be known as "thin-skinned" tectonics (Rodgers, 1949, 1950, 1963; Gwinn, 1964). J. L. Rich's (1934) concept of bedding-plane thrusts has proved to be an enormously powerful insight into the mechanics of the Appalachian Valley and Ridge Province and its adjacent plateaus to the northwest.

The existence of thrusts in the limestones of the "Little Mountains" belt has been recognized since the beginning of the present century (van Ingen and Clark, 1903). Various investigators have commented on these thrusts and have considered them to be a local phenomenon (Schuchert and Longwell, 1932; Chadwick and Kay, 1933; Chadwick, 1944; Dunn and Rickard, 1961). Some of these thrusts themselves have been folded.

In general the thrusts have not been studied seriously; at most they have been considered to be local breaks associated with ruptured fold limbs. The new roadcuts north of Austin Glen, to be visited on this excursion (Figs. 5 and 6), suggest that these thrusts are not local, isolated features, but on the contrary, are classic J. L. Rich-type bedding thrusts that have been folded. The astonishing features of these folded thrusts is not so much that they are present at all, but that in their westernmost exposures within the "Little Mountains," they appear not to rise to the surface. Instead they dip westward and evidently pass beneath the strata underlying the Catskills!

Generally it is not possible to determine how far these bedding thrusts have moved. My feeling is that the displacements are small--perhaps at most a fraction of a kilometer--but the evidence is ambiguous. But even if their displacements are small, the faults appear to be continuous and widespread and they lend striking support to the hypothesis that bedding thrusts underlie the entire Catskill-Pocono-Allegheny Plateau complex (Rodgers, 1963; Gwinn, 1964).

However one chooses to interpret the west dip of the bedding thrusts of the "Little Mountains," the very presence of such thrusts suggests diagnostic "Appalachian" orogenic products and, granting this, that the age of the deformation of the "Little Mountains," is more probably Appalachian than it is Acadian.

(b) Dying-out of folds. The "Little Mountains" remarkably parallel the Appalachian Valley and Ridge Province in that their folds die out where the Precambrian basement materials rise toward the present surface. The folds of the "Little Mountains" die out both northward and westward; the folds of the Appalachian Valley and Ridge Province die out northeastward. In both areas I think the cause is the same; where the rigid Precambrian basement stands high enough the thin covering strata were not folded. How high is "high enough"? The answer is not the same in the "Little Mountains" and in the Appalachian Valley and Ridge Province. By referring to the structure contours shown on the Basement Map of North America (Flawn, Chairman, and Kinney, Editor, 1967), one can acquire some idea of the inferred depth to basement at the places where the folds die out.

The folds of the "Little Mountains" die out northward at Clarksville, New York (Ruedemann, 1930). At this point, an estimated 2,000m of flat-lying strata cap the basement. In the Little Mountains, strata lacking notably competent members are folded where the basement lies 3,000m deep. At Kingston, where the competent Shawangunk Formation appears, the basement lies 4,000m deep. In northeastern Pennsylvania, where the large Appalchian folds die out, the depth to basement is approximately 6,000m, twice its depth in the "Little Mountains".

Could the configuration of the basement explain not only the dying-out of the folds but also part of the change of trend?

Similarity of regional joint systems

In both the "Little Mountains" and in parts of the Appalachian Plateaus the regional joint systems are similarly oriented and antedate the folds. This similarity was pointed out by Babcock (1966ms), but Babcock accepted the interpretation that the age of the folding of the "Little Mountains" was Acadian, so he did not discuss the possibility that the joints in both areas may have originated at about the same time. Although any argument based on joints must be tenuous if the age of the joints is unknown, nevertheless the similarities between joints and folds in both areas might be expected if the strata of both areas were deformed at the same time.

LOCALITY GUIDE

Proceed to Exit 21 of Gov. Thomas E. Dewey Thruway. After passing the toll gate turn **left** onto old N. Y. 23, 145 and drive southeast through Jefferson Heights to the junction of U. S. 9W. Turn N on 9W and then immediately right back onto N. Y. 23, 145. Stop 1 is the roadcut on the east side of the road across the street from several houses.

After everyone has had his fill of this exposure or 30 minutes, whichever comes first, return to the bus and proceed northwest on N. Y. 23-145 to Jefferson Heights. Turn southwest (left) into the short street marked Austin Acres. By the guard rail at the end of the street take the path that leads down into Austin Glen.

Persons who use this guidebook to go to Austin Glen on their own should request permission of the owner, Mrs. Helen Behrendt, telephone (518) 943-3813. Mrs. Behrendt lives in the old Austin homestead, the large stone house just northwest of Austin Acres. In the Glen be careful of fire.

No attempt has been made to prepare a stop-by-stop account of what is to be seen in the Glen and in Leeds Gorge, where we will complete the excursion.

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Map

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THE CATSKILL DELTAIC COMPLEX - DELTAIC PHASES AND CORRELATIONS OF THE MIDDLE DEVONIAN MARCELLUS FORMATION IN THE ALBANY REGION

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INTRODUCTION

While studying the clay mineralogy and trace element distribution of shales within the Marcellus Formation in southeastern New York for the purpose of paleoenvironmental analysis and interpretation, it became apparent that the persistence of certain "reworked" shoreline lithosomes could provide an additional basis for correlation and environmental interpretation (Wolff,1967a).

These rock units, occurring within the suggested time-stratigraphic units of Cooper (1933) and Goldring (1943) were found in the uppermost 150 feet of the Marcellus Formation between the Schoharie Valley and Port Jervis, New York (Figure 1). Their discovery led to the recognition of the contact between the Marcellus and Skaneateles Formations of the Hamilton Group; to the extension of the marine Marcellus subdivisions from central New York into the Catskill front; and to the further extension of both these features southwestward, 80 miles along the Hoogeberg escarpment,to the vicinity of Ellenville, New York (Wolff,1967b).

These units consist of a thin, sheet-like polymict, marine conglomerate and, somewhat higher in the section, a moderately-sorted, calcareous subgraywacke in the form of a series of small en-echelon barrier or offshore bars or beaches. Both units are quite thin (six inches for the conglomerate, six feet for the barrier bars), parallel to the outcrop and depositional strike, and appear to reflect the transgressive shoaling and strandline conditions characteristic of portions of the Upper Marcellus.

The marine conglomerate (Napanoch Conglomerate) has been traced from Ellenville to the southeastern corner of the Berne Quadrangle (Figure 1). Its westward extension as a time-stratigraphic unit to Schoharie Valley and central New York is based on the faunal zonations of previous investigators and the sedimentologic features noted within the upper Marcellus sequence. The bar-like, planar cross-bedded, sorted, calcareous subgraywacke (Mottville Sandstone), a unique lithologic unit that caps any vertical section of this formation, can be traced along the entire outcrop belt, but usually lies 5-25 feet above the last known marine beds. Its time-stratigraphic position is based on the association with a Mottville fauna in the Schoharie Valley and with the discovery of "Spirifer" sculptilis, a guide fossil for the Skaneateles Formation, by Goldring (1943) east of the Schoharie Valley in a marine re-entrant over 100 feet above the Mottville Sandstone.

Though outcrops are scattered and discontinuous, and faunal control for each section is still lacking, the use of the limited faunal criteria, in association with the recognition of distinct deltaic phases (and their vertical and lateral variations), seem to provide a better understanding of the tectonic framework and the relation between sedimentation and subsidence to the depositional environments. These associations can be applied, not only to the Marcellus Formation, or even the Hamilton Group, but to the lateral and vertical development of the entire Middle and Upper Devonian Catskill Deltaic Complex of New York.

TECTONIC CONTROL OF DEPOSITIONAL PHASES

During this study the writer attempted to provide a stratigraphic framework of neritic deltaic sedimentation for the classical "Catskill Delta" of New York (Wolff,1965). This was necessary because of the conflicting viewpoints regarding depth of water, the amount of depositional slope and the influence of tectonics, and inconsistencies within the general framework of facies and tectonic patterns outlined by the many previous investigators. The interpretations from this regional study are based on detailed observations within the lower Hamilton Group in southeastern New York, a reconnaissance survey of the Upper Devonian in central and south-central New York, the distribution of phases in the Devonian Correlation Chart (Rickard,1964), and a survey of the literature.

Regional Scale

Because of the external geometry of many ancient deltas and the lack of impress of the deltaic environments on the sediments Krumbein and Sloss (1963) refer to these features as clastic wedges. The reason for this term is due to the tectonic influence on the sediments because: 1) the dominantly subsiding tendencies of the marginal basin cause sediment burial before the environment can influence the composition and geometry of the sediments; 2) the variable strandline position is determined by the rate of sediment supply which reflects the variable tectonic activity of the source area; 3) the borders of the sea are subject to continuous large scale advances and retreats with a short duration of the deltaic environment at any one point.

Perhaps none of these points are applicable to the Devonian Catskill Complex because Krumbein and Sloss fail to distinguish between the constructional and destructional phases of delta building (Scruton,1960), and each of the environments may have exerted some control on the character of the sediments, regardless of the tectonic influence. Though the Catskill Delta is almost entirely constructional (as reflected by the sediment phases), there are preserved a few short durations of the destructive aspects which allow sorting and reworking of previously deposited material. Perhaps the term "tectonic delta complex" (Friedman & Johnson,1966) is more appropriate. It is noteworthy that during any of the destructional phases the distal parts of the basin tend to become areas of limestone accumulation.

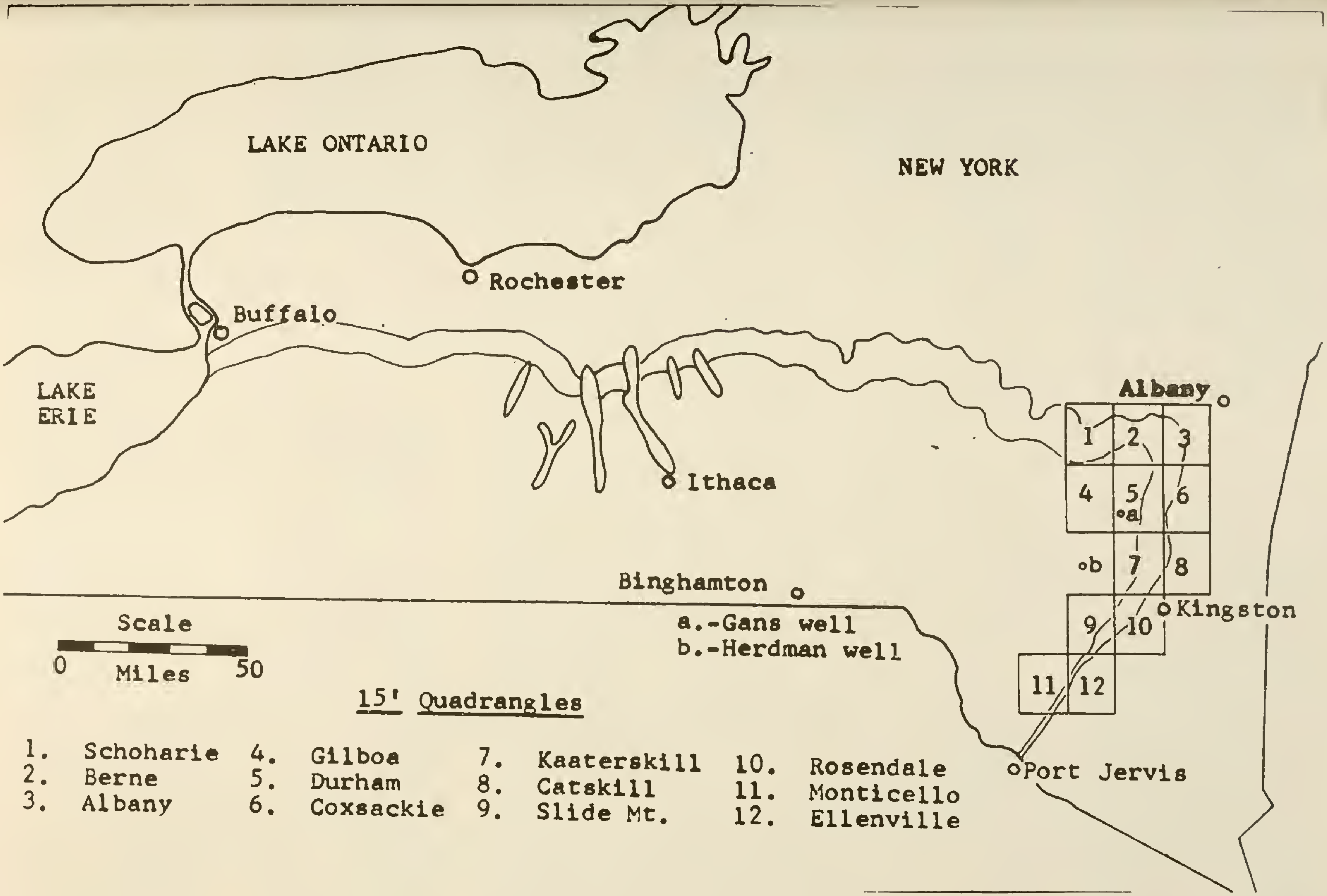

Figure 1. Location of area of detailed study of the Marcellus Formation and its subdivisions

Local Scale

One of the chief lines of evidence used to distinguish the basin, slope, and shelf environments of the Upper Devonian in central and western New York is the positioning of the turbudite sequences of the Chagrin phase (Rickard,1964) across the region (Rich,1951, Sutton,1960,1963, McIver, 1961). Since many modern analogs of turbidite sequences occur on continental slopes at bathyal depths, few could consider the Upper Devonian marine section as being deltaic in origin (Colton & de Witt, 1958, Pepper & de Witt, 1950,1951, Pepper, de Witt & Colton, 1956).

It seems that, if one could provide a mechanism for the origin, transport and deposition of the density currents on very gentle slopes, there would no longer be a need to envision steep slopes and bathyal depths in the environmental interpretation, and the basin-slope-shelf descriptions would be more applicable to deltas. The mechanism suggested (Wolff,1967a) involves the use of seismic vibrations and earthquakes as the triggering and transporting mechanism that preceeded or accompanied the orogenic folding that took place in the Appalachian basin during deposition of the Upper Devonian strata.

The concept of contemporaneous folding and sedimentation initiated by Willis (1893), has been re-examined and popularized by Cooper (1961), and has been noted in the Upper Devonian of New York by Woodrow (1964), and in the subsurface by Bradley & Pepper (1938). Recent investigations (Fletcher & Woodrow,1967) extend this orogenic interval to the Middle Devonian strata. The general orientation of the sole markings is to the west and southwest (Sheldon, 1929, Sutton, 1959, McIver,1961, Colton, 1967) down the paleoslope and parallel to the "growing" buried major structures in the basin. This alignment suggests that the earthquake zones, with epicenters parallel to these major folds and faults, could activate a series of density currents at different elevations on the paleoslope, and thus eliminate the necessity of steep slopes in the marine portion of the Upper Devonian Catskill Delta Complex.

This mechanism has been used by Sorauf (1965) to account for the origin of flow rolls or "ball and pillow" structures (Potter & Pettijohn, 1963) that occur so frequently in the Smethport phase (Rickard, 1964) near the transition between shelf and slope deposits. These large nodules of soft-sediment deformation may have begun as broad marine shelves or sand channels that were later contorted and foundered by vertical movement induced by seismic vibrations.

SEDIMENTOLOGIC DESIGN OF DELTAIC SEQUENCES

Provenance and Dispersal Patterns

A summary of the work related to major sediment patterns in the Appalachian basin was outlined by Pryor (1961) who indicated that the westward transport direction, paleoslope and facies gradients are normal to the nearly north-south depositional strike while the major sand body trends are parallel to it. Other workers (Pelletier, 1958, McIver,1961, Burtner,1964) indicate that isopachs, sandstone/shale ratios, contours of maximum pebble size, and lines of equal bedding thickness are also parallel to the depositional strike. This information, in conjunction with the paleocurrent and petrographic data (Menscher, 1939, Pelletier, 1958, McIver, 1961, Burtner, 1964, Fletcher, 1964, Buttner, 1965, 1968, Lucier, 1966) substantiate the linear source to the northeast, east, and southeast as originally postualted by Barrell (1913, 1914).

Menscher (1939) was the first to demonstrate that the source for the northern Catskills could have been the Taconic Mts. of New England while the sediments in the Skunnemunk outlier, 40 miles southeast of the Catskill Mts. may have come from the Manhattan Prong. This has been substantiated by Burtner (1963) and Lucier (1966) who found evidence suggesting that the eroded Ordovician and Silurian sediments as well as the low-rank metamorphics of the New England Taconics provided a source for sediments in the northern Catskills while strata in Pennsylvania were derived from the granites and gneisses of the Manhattan Prong.

Characteristics of Devonian Depositional Phases

Most of the early work on the Devonian strata in New York and adjacent states was concerned with methods of correlation between different depositional phases and the significant and distinctive lithologic features associated with each. More recent contributions have been more elaborate and extensive (Chadwick, 1933, 1935, Grossman, 1944, Pepper, et. al., 1950, 1951, 1954, 1956, Colten & deWitt, 1958, deWitt, 1960, Sutton, 1960, 1963, McIver, 1961, Rickard, 1964, Wolff, 1965, Friedman & Johnson, 1966, Buttner, 1968, McCave, 1968, 1969). From these investigations has arisen the recognition of repeating rhythmic depositional phases (Caster, 1934, Rich, 1951, Rickard, 1964) which are here interpreted as ancient deltaic environments of tectonic origin (Wolff, 1965). The main characteristics of each phase have been described by nearly all these investigators and have recently been summarized by Rickard, (1964). Only certain aspects are presented here to illustrate their relation to modern deltaic environments (Figure 2).

1. Cleveland Phase - consists of relatively thin, laminated, fissile black shales, often calcareous or pyritiferous. It contains a sparce pelagic fauna and sedimentologic features characteristic of material being slowly deposited from suspension in an isolated euxinic marine basin. It is the fondoform of Rich (1951), the basin of Sutton (1963), and the delta toe in this study (Figure 2). Modern analogs could include the offshore marine clays of the Mississippi (Scruton, 1960), Niger (Allen, 1965) or several other deltas (Shirley & Ragsdale, 1967).

2. Chagrin Phase - contains thin-bedded, dark gray fissile shales interbedded with lesser amounts of thin, laterally persistent, even-bedded or laminated siltstones or fine-grained sandstones. The coarser units may show graded bedding and usually exhibit sharp basal contacts that become gradational with the overlying shale. They may also contain the plane-bed, laminated, and then micro-cross-bedded structures through a thin vertical sequence - all the characteristic features of turbidite deposits. Other features

include the cyclical nature of these deposits, the persistent regional orientation of current structures, the associated sole marks and load casts, and the small-scale ripples and cross-lamination. Fossils are scarce and consist chiefly of transported mollusks and brachiopods as local coquinites near the base of siltstones. This environment is the clinoform of Rich (1951), and the slope of Sutton (1963), but because of the large areal extent of these turbidite sequences, it appears to be transitional between the major portion of the prodelta slope and the delta toe deposits, and is referred to as the distal prodelta slope in this study (Figure 2). Modern anologs include the prodelta slope of recent deltas, but for most there is little indication of large scale deposition from the influence of density currents.

3. Big Bend Phase - characterized by the presence of thin-bedded and blocky arenaceous shales, by the general absence of laminations, and by the development of upward-coarsening cycles of regional extent. This phase usually forms the thickest part of any contemporaneous regressive marine sequence. Though shales dominate, interbedded siltstones and sandstones become more common and more massive in the upper portion of the phase. In the Upper Devonian persistent horizontal laminations, flute casts, and lobate flow markings are common, and numerous turbidite sequences, similar to those of the Chagrin phase, make up much of the section. But these features are rare in the Hamilton Group and most of the marine sediments are within the Big Bend phase of sedimentation. This phase is not well-developed in the Upper Devonian clastics - and thus there is a frequent reference of a transition from the Chagrin to the Smethport phases of deposition (McIver,1961, Sutton,1963).

The interbedded sandstones, while becoming more massive, usually exhibit no visible textural gradations and, while normally persistent, may pinch-out or divide into sets of beds higher in the section. Local lensing and broad, shallow channeling may occur in the transition to the overlying Smethport phase. The sediments are characteristic of the prodelta slope of recent deltaic environments and are referred to as the proximal delta slope in this study (Figure 2).

4. Smethport Phase - the environment of maximum variability in sediment type, associated structures, and sediment distribution. It consists largely of even or irregularly bedded olive subgraywackes and siltstones with lesser amounts of olive and gray-dark gray shales and silty mudstones. It is the first phase in which horizons of shale -clasts, plant fragments, and conglomerates may be noted. Turbidite associations are not common, but "ball and pillow" structures are abundant. Horizontal or shallow simple and planar cross-bedding and primary current lineation are common in the gray subgraywacke flagstones (most are of beach, barrier bar, or point bar origin). Cross-bedding sizes and types are variable and most occur as deposits from slackening currents in intertidal or alluvial channels. The thin units show a variable orientation and seem to be associated with nearshore marine or lagoonal sequences. The medium and thick-bedded units indicate a more consistent west and northwest orientation, numerous erosion surfaces with abundant intraformational conglomerates and plant fragments, and seem to be associated with

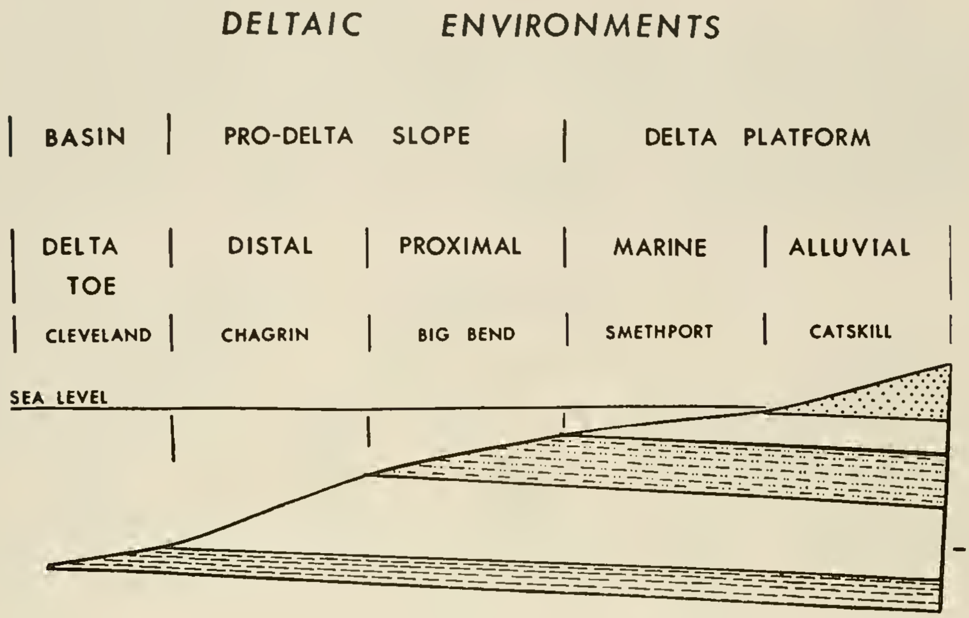

Figure 2. Relation of Devonian sediment phase (facies) to deltaic environment

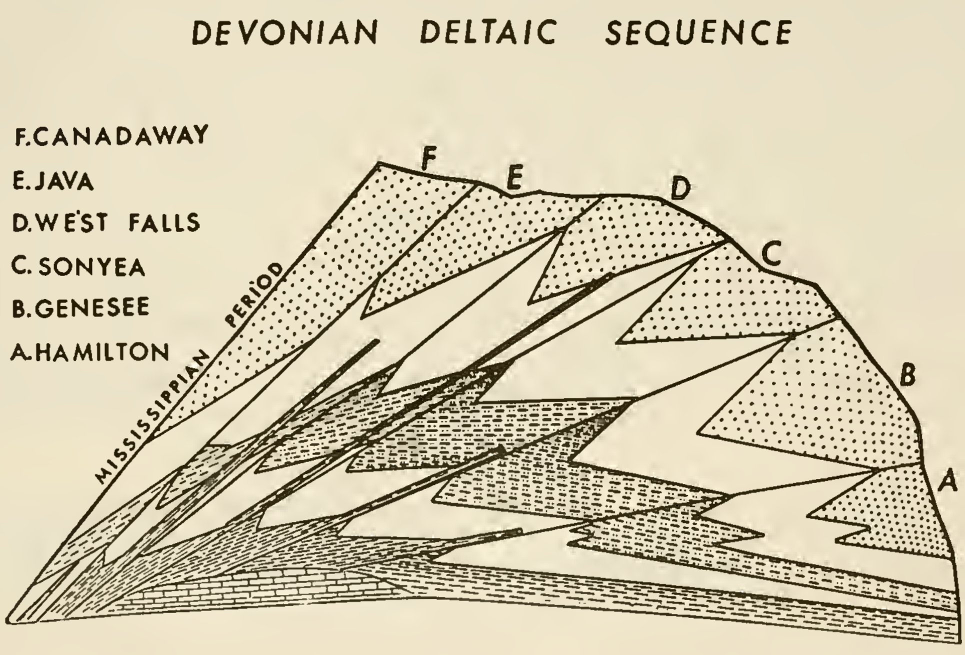

Figure 3. Vertical sequencing of Middle and Upper Devonian deltaic deposits into 6 deltas based on the black shale extensions

broad tidal channels.

Upward coarsening cycles are frequent, and still characteristic, but upward-fining cycles with numerous broad erosional channels indicate lateral as well as vertical accretion and suggest the mechanism which accounts for the rapid lithologic changes. Fossil assembleges occur chiefly as coquinites associated with massive fine-grained sandstones, flagstones, or thin cross-bedded subgraywackes, and are dominated by spiriferoid brachiopods and pelecypods. Conglomerate horizons are also lenticular (surf zone reworking) or may occur as distinct sheets of intraformational shale or limestone fragments.

This environment has been described as undaform by Rich (1951), shelf by Sutton (1963) and is here considered as the marine delta platform (Figure 2). Modern analogs include the delta front platform of recent deltas. Other environments within this phase may include: interdistributary bays, distributary mouth bars, barrier bars, offshore bars, cheniers, tidal channels, tidal flats and lagoons, beaches, marshes, and spits. Several investigators have listed criteria used to differentiate the deposits characteristic of each of these environments. A partial list includes: Fisk, et. al., (1954), Scruton (1960), Shepard & Moore (1955), Shepard (1960), Lane (1963), Allen (1965), Visher (1965), Potter (1967), Hoyt (1967, 1969). In many instances the criteria used to characterize the sediments associated with the Mottville Sandstone are based on these references.

5. <u>Catskill Phase</u> - composed predominantly of cyclothems consisting of large scale planar and trough cross-bedded, gray and red subgraywacke and micaceous sandstones with lesser amounts of red mudstones, siltstone and shale, and small quantities of olive mudstones, dark shales and quartz arenites. Quartz and shale pebble conglomerates and plant fragments are locally abundant. These upward-fining cyclothems and the presence of redbeds are the characteristic features, gray-green sandstones usually making up 60-80 percent of each cycle.

Local and regional effects of compaction and subsidence are reflected in the development of rhythmic sequences which can be separated into transgressive, stable, and regressive phases (Buttner, 1966). The inequalities in the balance between subsidence and sedimentation account for the lateral distribution of phases and for the vertical changes of position between major distributaries and embayments.

In the Hamilton Group vertical sequences usually consist of an upward fining cycle changing from gray subgraywacke and siltstone on an erosional base that may contain shale fragments and cut and fill structures, to blocky red and olive mudstones and shales, frequently cut into or across by the next overlying channel sandstones. More complete rhythms have been described by Buttner (1968) for the

Upper Devonian which includes polymict conglomerates at the base and dark shales and well-sorted subprotoquartzites at the top. The presence of abundant conglomerates is characteristic of the "Pocono" phase of sedimentation.

Channels are defined by planar and trough cross-bedded graywackes and subgraywackes bounded by irregular erosion surfaces. Most sets are lenticular with one to two inch cross-laminations at angles of 3-10 degrees. Thick planar sets are more common than the trough types in the initial sequence of redbeds and appear to indicate point bar accretion during lateral migration of stream or tidal channels. Rare megaripples indicate the sediment transport transition from lower to higher flow regimes and the migration of these forms into large scale cross-bedding.

The red siltstones and mudstones and olive mudstones form the floodplain deposits, but unlike recent deposits of similar origin, appear to lack thin horizontal bedding and lamination. The red units are thickest near the channel deposits and thin laterally away from them as they grade into olive mudstones. Small tributary channels of thin cross-bedded sandstones may cut through some of these deposits and erode portions of the olive or red mudstones near the center of the floodplain.

The occurrence of dark gray shales with an associated brackish fauna (Buttner, 1965, McCave, 1966) and light gray subgraywackes indicate that certain areas of depression between major distributary systems may have been opened to the sea as lagoonal or interdistributary bay environments during periods of maximum subsidence. The lack of preservation of these deposits is due to the frequent shifting and erosion of aggrading river channels across these regions. The Catskill phase has been defined as being subaerial or fluvial and is considered as the alluvial delta platform in this study (Figure 2).

Effects of Subsidence on the Depositional Phases

If the sediments phases were affected by earthquakes it becomes possible to consider the entire sedimentologic framework of the Devonian "tectonic delta complex" as being controlled by the almost continuous progradation of clastic material into relatively shallow water during nearly continuous but differential basin subsidence, which led to the development of a series of overlapping rhythmic deltaic sequences (Wolff, 1965).

While the Middle Devonian shows the initial aspects of the Acadian orogeny and is characterized by a single (Hamilton) delta with periods of limestone deposition, the Upper Devonian exhibits much greater tectonic activity with nearly continuous sedimentation, and a series of overlapping rhythmic deltaic deposits are formed. Each interval begins with a period predominated by subsidence and a barring of the basin on the east side of the Findley Arch, producing a thin but a really extensive black shale (delta toe environment). Due to the subsidence, this unit forms over a previously deposited gray shale and siltstone (distal delta slope) and there is a landward shift of all the deltaic environment (Rickard, 1964) and then the gradual build-out of a new wedge (delta) of marine and continental sediments (Figure 3). Thus, each delta extends from the base of one black shale to the base of the next, and its name follows the name of the associated stratigraphic unit - Genesee, Sonyea, West Falls, Java, and Canadaway (Figure 3).

The sediment phases of the Devonian Catskill tectonic delta complex of New York (Rickard, 1964) can be related to different facies of modern deltaic environments (Shirley & Regsdale, 1967). Each phase (except Chagrin) contains features usually associated with shallow water (neritic) depositional environments. Gentle prodelta slopes and shallow distal marine environments are also suggested by the widespread development of black shales across several sediment phases and by the numerous localized domes and basins that affected the geometry and distribution of associated sediments in the deeper parts of the basin.

REGIONAL STRATIGRAPHY

Historical Development

The Devonian of New York has been recognized as a classic example of deltaic sedimentation in a subsiding basin since the work of Barrell (1913, 1914). The earliest studies of the Hamilton Group in southeastern New York are found in the reports of the first state geological survey by Mather (1840) and Vanuxem (1842). The Marcellus Formation was not included in the Hamilton Group until the work of Darton (1894). It remained for Cooper (1933) and Goldring (1935, 1943) to subdivide the Marcellus and to locate the top of the marine Marcellus east of Schoharie Valley. The boundary for the top of the Marcellus within the non-marine strata and the extension of the Marcellus subdivisions into the Catskill Mountains and southwestward along the Hoogeberg escarpment was suggested by Wolff (1967_b).

Marcellus Correlations

The subdivisions of members proposed by Cooper (1933) could not be extended east of Schoharie Valley due to the lack of guide fossils and the gradual facies changes; the last horizons of marine fossils were known to occur near the top of the Marcellus Formation. In this region the lower part of the Mottville Sandstone contains *Paraspirifer acuminatus* and *Tropidoleptus sp.* Above this interval is the first occurrence of "*Spirifer*" *sculptilis*, an index fossil for the Skaneateles Formation (Rickard, 1964).

The lithology between these time-stratigraphic markers gradually changes from marine to continental but does include the calcareous subgraywacke known as the Mottville Sandstone. The top of the Marcellus east of the Schoharie Valley is therefore designated as the interval just below the last appearance of Paraspirifer acuminatus and the first appearance of "Spirifer" sculptilis. Since Sp. sculptilis has only been found in one locality east of Schoharie Valley, a more practical designation would be to include the interval below the last appearance of Paraspirifer acuminatus and below the first appearance of the bar-like subgraywacke referred to as the Mottville Sandstone. The suggested correlations for the Marcellus Formation and its subdivisions are illustrated in Figure 4.

The thickness of the Marcellus is estimated by Cooper (1933) to be 800 feet near Richmondville, and 900 feet in Schoharie Valley. Using the interval between the Mottville Sandstone and the Onondaga Limestone as measured in the composite sections and estimated from the projected thickening rates from the Devonian Correlation Chart the writer would add: about 1,200 feet in the Alcove Reservoir area (near northeastern end of basin axis); 1,000 feet at Potic Mt.,; 1,200 feet on Route 28 east of Ashokan Reservoir; and, continuing south, about 1,400 feet near Napanoch, New York.

Description and Correlations of Marcellus Subdivisions

Since outcrops are of limited extent, a series of composite stratigraphic sections based on elevations and dip measurements have been included (Figure 5). A more complete description of each exposure along with the fossils, clay mineralogy and trace element distribution is provided by Wolff (1967_a). Some of the paleoenvironmental interpretations, based on current information, have been updated since that time. A list of important fossils with several illustrations is given by Dunn and Rickard (1961).

1. Union Springs Member - known locally as the Marcellus or Bakoven black shale (Chadwick, 1944). It is a carbonaceous, soft, thin-bedded, fissile shale representative of the "Cleveland" depositional phase. Overlying the Onondaga, it is in turn overlain by the Cherry Valley Limestone west of Schoharie Valley, and the laterally equivalent Stony Hollow member east of the valley and below the Catskill front (Figures 5 & 6). The basal contact with the Onondaga Limestone is quite abrupt and Chadwick (1927) recorded an unconformity in the Catskill region that has also been noted at Mill Hook, southwest of Kingston. Cooper (1930) realized this was a marginal break that does not extend into the basin. The upper contact is frequently marked by strong shearing and deformation, both in outcrop and well samples, and has been noted by Rickard (1952) near the type section of the Cherry Valley Limestone - a distance of nearly 40 miles from the Catskill front.

2. Stony Hollow Member - the silty equivalent of the massive, gray Cherry Valley Limestone of western New York. As it becomes more shaley and silty it passes into the "Moscow" phase of sedimentation (Rickard, 1964). It consists of repetitive cycles of interbedded shales and calcareous siltstones or fine-grained

<table>
<tr><td rowspan="2">TIME - ROCK UNIT</td><td colspan="7">STRATIGRAPHIC CORRELATIONS</td></tr>
<tr><td colspan="2">Otsego Lake</td><td colspan="2">Unadilla Valley</td><td>Schoharie Valley</td><td>Northern Helderbergs</td><td>Hoogeberg Escarpment</td></tr>
<tr><td></td><td colspan="7">PREVIOUS CORRELATIONS</td></tr>
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<tr><td></td><td colspan="7">PRESENT CORRELATIONS</td></tr>
<tr><td rowspan="2">SKANEATELES FORMATION</td><td colspan="2">Delphi Station</td><td colspan="2">Panther Mt.</td><td>Panther Mt.</td><td>Ashokan</td><td>Plattekill</td></tr>
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<tr><td colspan="4">Cherry Valley</td><td colspan="3">Stony Hollow</td></tr>
<tr><td colspan="4">Union Springs</td><td colspan="3">Union Springs</td></tr>
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Figure 4. Previous and present Marcellus correlations from central into southeastern New York

sandstones but does not show other characteristic features usually associated with turbidites. The member is 25 feet thick in the Albany Quadrangle and increases to 140 feet at the type section west of Kingston.

The basal contact is a transitional sequence of medium-grained calcareous sandstones interbedded with black shale and some crystalline limestone as a fracture filling. This zone is marked by some deformation and vertical cleavage indicating the residual effects of the overthrusting below. The upper contact with the Chittenango Member (lower Mt. Marion of Chadwick, 1944) is also transitional and marked by change into black, non-calcareous, massive arenaceous shales with a blocky or knobby fracture pattern, and a few thin siltstones.

3. Chittenango Member - a continuation of this unit from central New York (Figure 6) is based on thickness of strata between the Stony Hollow and the *Meristella* - coral horizon. East of Schoharie Valley and in the Berne Quadrangle the black fissile shales of the Chittenango (Cleveland Phase) are replaced by arenaceous shales and siltstones of the "Big Bend" phase with the appearance of the coral horizon near this contact. Since the thickness of the characteristic Chittenango lithology decreases as the section becomes more arenaceous eastward, nearly 100 feet of the Stony Hollow must be temporally equivalent to this unit. An additional 120 feet is supplied by the section between the top of the Stony Hollow and the *Meristella* - coral horizon. Thus the Cleveland phase of the Chittenango west of Schoharie Valley changes to the Moscow and Big Bend phases as it increases in thickness eastward.

4. Otsego Member - this and the overlying Solsville Member make up a large part of the previously undivided Mt. Mation Formation of Chadwick (1944) which forms the prominent portion of the Hoogeberg escarpment beneath the Catskill Mts. (Figures 5 & 6). As also indicated by Rickard and Zenger (1964) in the Cooperstown Quadrangle, both these units consist of upward coarsening sequences.

The Otsego lithology is similar to the Big Bend phase, consisting of dark gray fissile and arenaceous shales with light gray lenses and layers of siltstones and fine-grained sandstones that gain prominence as one ascends the section. The turbidite features associated with the Chagrin phase are lacking and there are no fossil horizons available for zonation. Along the Hoogeberg escarpment, the lower and upper portions of the Otsego can be distinguished by the distribution and type of siliceous concretions that occur in the shales.

While the lower contact of the Otsego has a time-stratigraphic basis, the top of the unit was mapped by lithology and a general faunal assemblege. Cooper and Goldring (in Goldring, 1943), on the basis of this assemblege, located the upper part of the Otsego in the Berne Quadrangle and the top of the Otsego in the Coxsackie and Catskill Quadrangles. In both areas the top of the Otsego is

STRATIGRAPHIC SECTIONS OF MARCELLUS FORMATION

IN SOUTHEASTERN NEW YORK

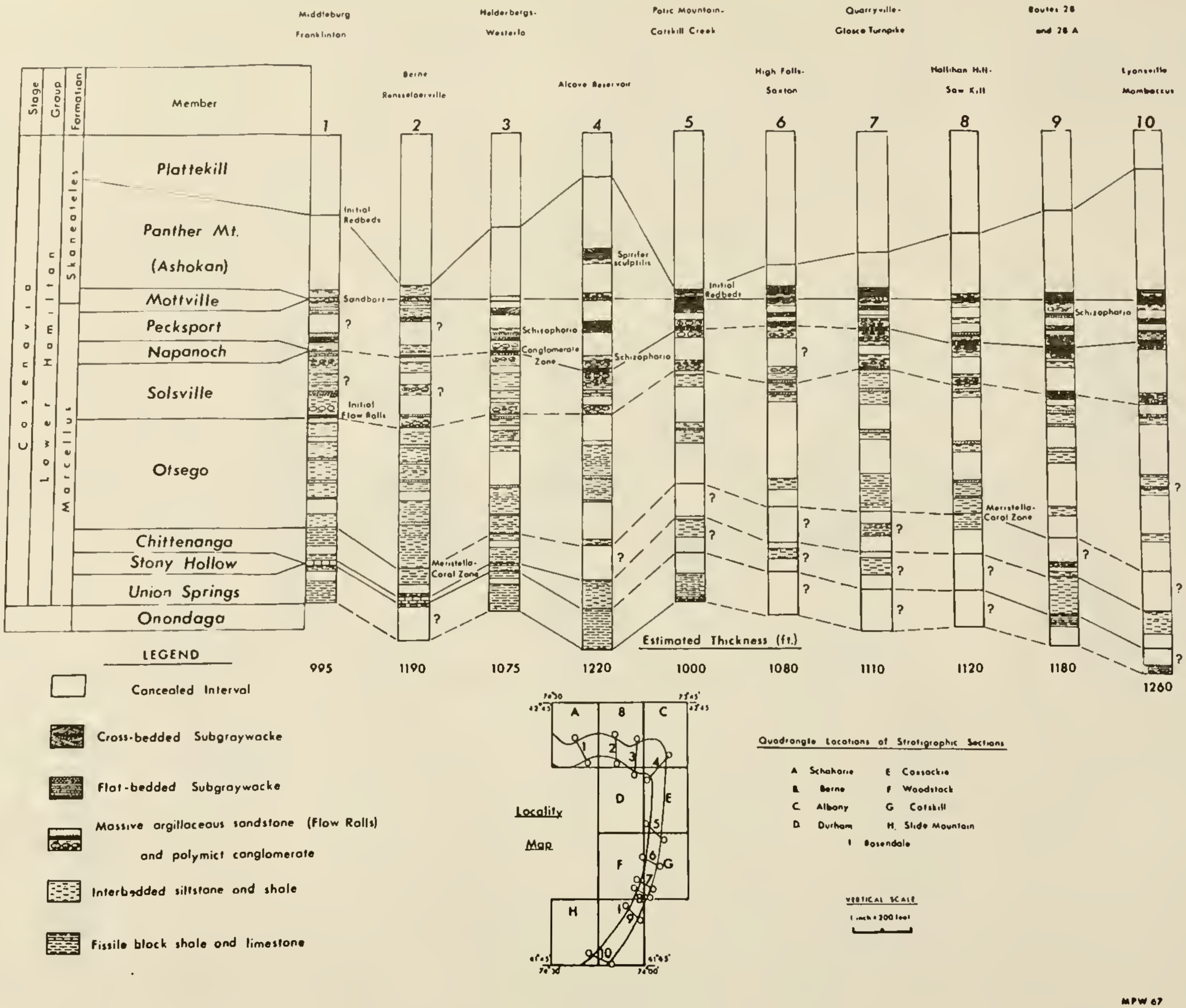


Figure 5.

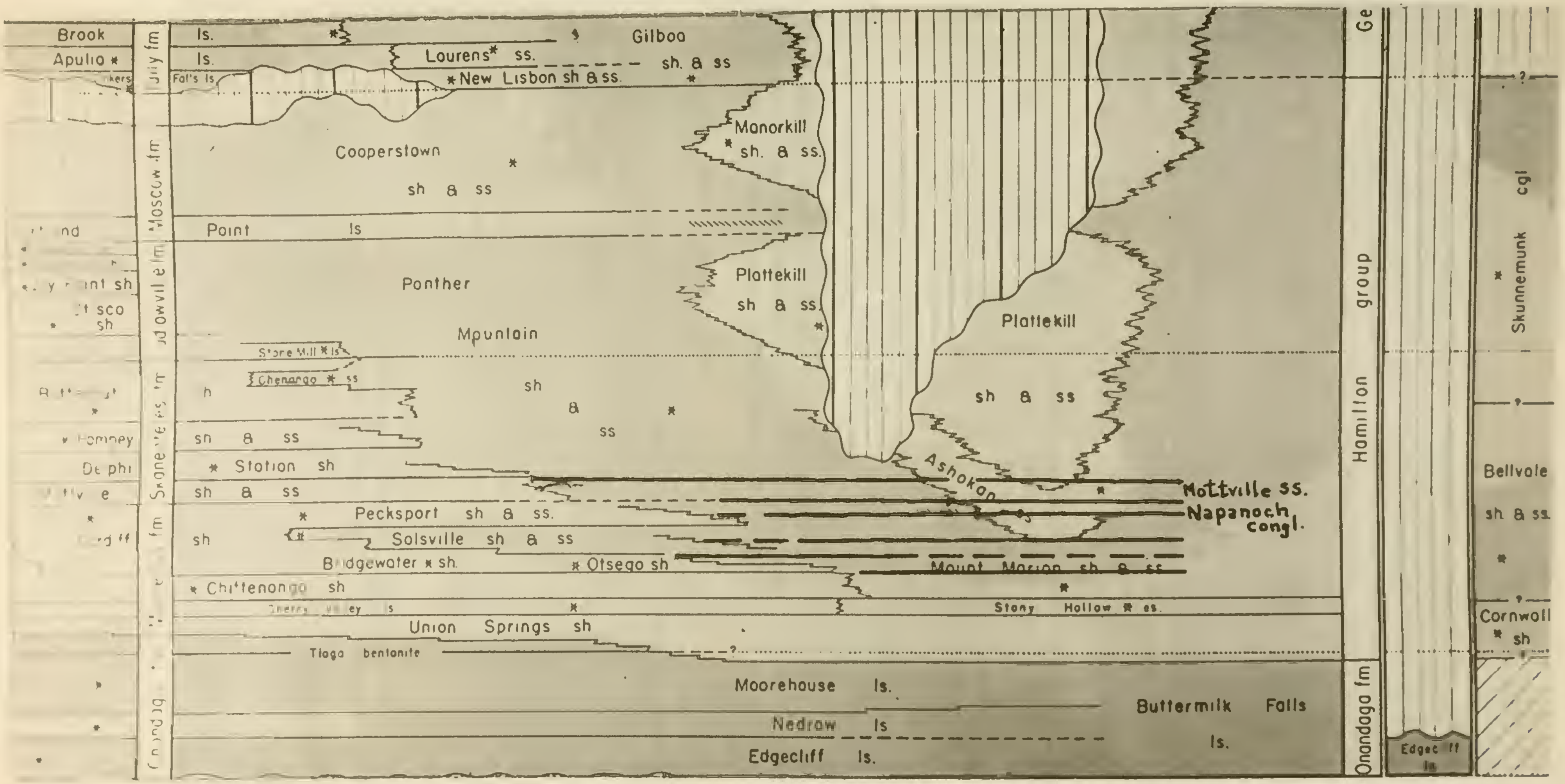


Figure 6. Proposed Marcellus correlations on the Devonian Correlation Chart

capped (within 20-30 feet) by a zone of "ball and pillow" structures (flow rolls) or massive and laminated fine-grained sandstones. This lithologic contrast has been found to be laterally persistent along the Hoogeberg escarpment and seems to be a valid criterion for noting the top of the Otsego or base of Solsville Member.

5. Solsville, Pecksport, and Mottville Members - the lithologic criteria and faunal assembleges defining these units vary to some degree from Schoharie Valley to the Hoogeberg escarpment. Since their total stratigraphic interval is not excessively thick and the section is capped by a rock and time-stratigraphic marker, an attempt has been made to subdivide these units whenever possible across each of the quadrangles (Figures 5 & 6).

Schoharie Quadrangle

Cooper noted that east of Cooperstown the massive, fine-grained Solsville continued to persist and separate the Otsego and Pecksport shales. In Unadilla Valley and eastward, this was no longer the case, there was also a gradual faunal replacement, and the entire interval was referred to as the Panther Mt. Formation (Figure 6).

In the Schoharie Valley, Cooper extended the top of the Otsego to the base of Towpath Mt., and to Bouck's Falls, on Panther Creek, south of Towpath Mt. The presence of fine-grained sandstones and "ball and pillow" structures capping the shale sequence of the Otsego indicates the continuation of an upward coarsening sequence (as also noted by Rickard and Zenger) near the base of the Solsville.

On the east side of Schoharie Valley the lower third of Walhalla Mt. contains a fauna (described by Prosser, 1899) and lithology suggestive of the criteria used to define the upper Otsego along the Hoogeberg escarpment. The upper portion of this sequence, near Breakabeen, has a fauna and lithology similar to the gorge above Bouck's Falls and suggestive of the upper Solsville. Due to lack of outcrop, differentiation of the Solsville and Pecksport was not attempted, but the Mottville (highest occurrence of *Paraspirifer acuminatus*) was located by Cooper at Hony Hill and this region also contains the calcareous subgraywackes characteristic of the Mottville Sandstone, (best exposed on Route 145 north of Franklinton).

Marine strata (Smethport phase) occur in this region for another 700 feet (Panther Mt. Member of the Skaneateles Formation) but these grade rapidly into the redbeds that establish the base of the time-transgressive Plattekill Member (Catskill phase).

Berne Quadrangle

The upward coarsening sequence of arenaceous shales into massive fine-grained subgraywackes and thin-bedded flagstones, characteristic of the Solsville farther to the east first appears in Brat Hollow southwest of Berne. Higher in the section Prosser (1899) has described a fauna suggestive of the top of the Pecksport that has also been noted by Goldring in other parts of the quadrangle. The strata are capped by calcareous greenish subgraywackes (Mottville Sandstone); redbeds occur forty feet above this interval (Figure 5).

At Rensselaerville the Solsville and Pecksport could not be faunally or lithologically differentiated, but some small barrier bars suggest the presence of shoaling that is characteristic of the Mottville. The associated olive and gray shales contain a brackish fauna (Goldring, 1935) and are overlain by red and green blocky mudstones.

The fine-grained sandstones and "ball and pillow" structures defining the base of the Solsville occur in South Berne. The presence of Schizophoria striatula and several other fossils is indicative of the Pecksport. Several coquinite horizons in "flagstone" quarries containing this assemblege were located near Westerlo, near the southern border of the Berne Quadrangle, and farther south, in the Durham Quadrangle.

A similar fauna (without Schizophoria) again as coquinite lenses occurs northeast of Westerlo, 150 feet lower in the section. This locality defines the top of the Solsville, and has also been found 20 feet above a marine conglomerate in Dormansville. This is the first location of the Napanoch Conglomerate, a key zone used to define the top of the Solsville and base of the Pecksport (Figure 7) from this region 75 miles southwestward to Ellenville, New York.

Coxsackie Quadrangle

The contact between the Otsego and Solsville Members, when faunal evidence is lacking, has been taken at the initial position of massive fine-grained sandstones above a thick sequence of dark arenaceous shales. Outcrops are few along the northern part of this region, but the contact is projected by the marked sharpening of relief at 1,500 feet at Wolf Hill in the Northern Helderbergs, 1,250 at Cass Hill, 1,130 on Derbyshire Road, and 900 feet on the hill northwest of Alcove Reservoir. The top of the Otsego was time-stratigraphically located by Cooper (1933) 3 miles south in the Alcove Reservoir Spillway at an elevation of 500 feet. "Ball and pillow" structures with coquinites occur 100 feet higher, and the section is capped by spheroidal quartz and rod-shaped siltstone and chert pebbles in a sandy matrix also containing macerated remains of Mucrospirifer (Figure 5).

A calcareous subgraywacke, weathering reddish, overlies 15 feet of arenaceous shales and mudstones on the Alcove-Newrys Road, 240 feet higher in the sequence. Coquinites in "flagstones" occur 140 feet higher in the section. This area, first studied by Goldring (1943), is important because:

1) It is the only area known, east of the Schoharie Valley,

where a marine fauna occurs above the sandstones representing the Mottville Member.

2) The faunal assembleqe is distinguished by the presence of "Spirifer" sculptilis (index fossil for the Skaneateles Formation) and Pterinopecten macrodonta, last noted 10 feet below the top of the marine Marcellus at Rensselaerville Falls. Also important is the absence of Paraspirifer acuminatus and Schizophoria striatula, last seen in the Pecksport Member.

Correlations in the southern part of the quadrangle are rock-stratigraphic. The Napanoch Conglomerate has been noted at Cole Hollow and on Catskill Creek near the south end of Potic Mt. The Mottville Sandstone can be located at the Spillway east of Potic Reservoir and in Catskill Creek directly below the first redbeds.

Catskill & Kaaterskill Quadrangles

From the vicinity of Potic Mt. near the northern border of the Catskill Quadrangle the Otsego and Solsville Members (Mt. Marion of Chadwick, 1944) extend as a continuous ridge of arenaceous shales and massive, dense, siltstones and thin cross-bedded subgraywackes (the Hoogeberg escarpment) that extends southwestward to Port Jervis. Outcrops can be examined along the numerous streams and roads that dissect the ridge at various intervals and provide a moderate degree of stratigraphic control between sections (Figure 5). The fauna duplicates that of the northwestern sections and a complete list has been provided by Prosser (1899), Goldring (1943), and Chadwick (1944).

The section at High Falls on the Kaaterskill includes the Solsville and Pecksport, but the conglomerate has not been located here - though it does occur north and south of this locality. The Mottville Sandstone occurs 0.4 miles to the west. The most complete section occurs along Route 32 near Quarryville where a nearly continuous section between the Stony Hollow and the Panther Mt. is located. Another good section is found between Hallihan Hill and Sawkill Creek. The Mottville Sandstone occurs 0.4 miles westward on the Zena-Saw Kill Road south of Kingston Reservoir.

Rosendale and Slide Mt. Quadrangles

The Solsville Member occurs near the top of a quarry on the south side of Route 28 (west of Stony Hollow) while "ball and pillow" structures and the Napanoch Conglomerate occur in road cuts on the north side of Route 28 about 150 feet higher. The top of the Pecksport (with Schizophoria striatula) occurs in a quarry on Route 28, 0.3 miles west of Binnewater Lake. The conglomerate and the Mottville Sandstone are well exposed on Route 28 and 28A between Stony Hollow and West Hurley. Further sections continue to the south, but are not pertinent to this field trip.

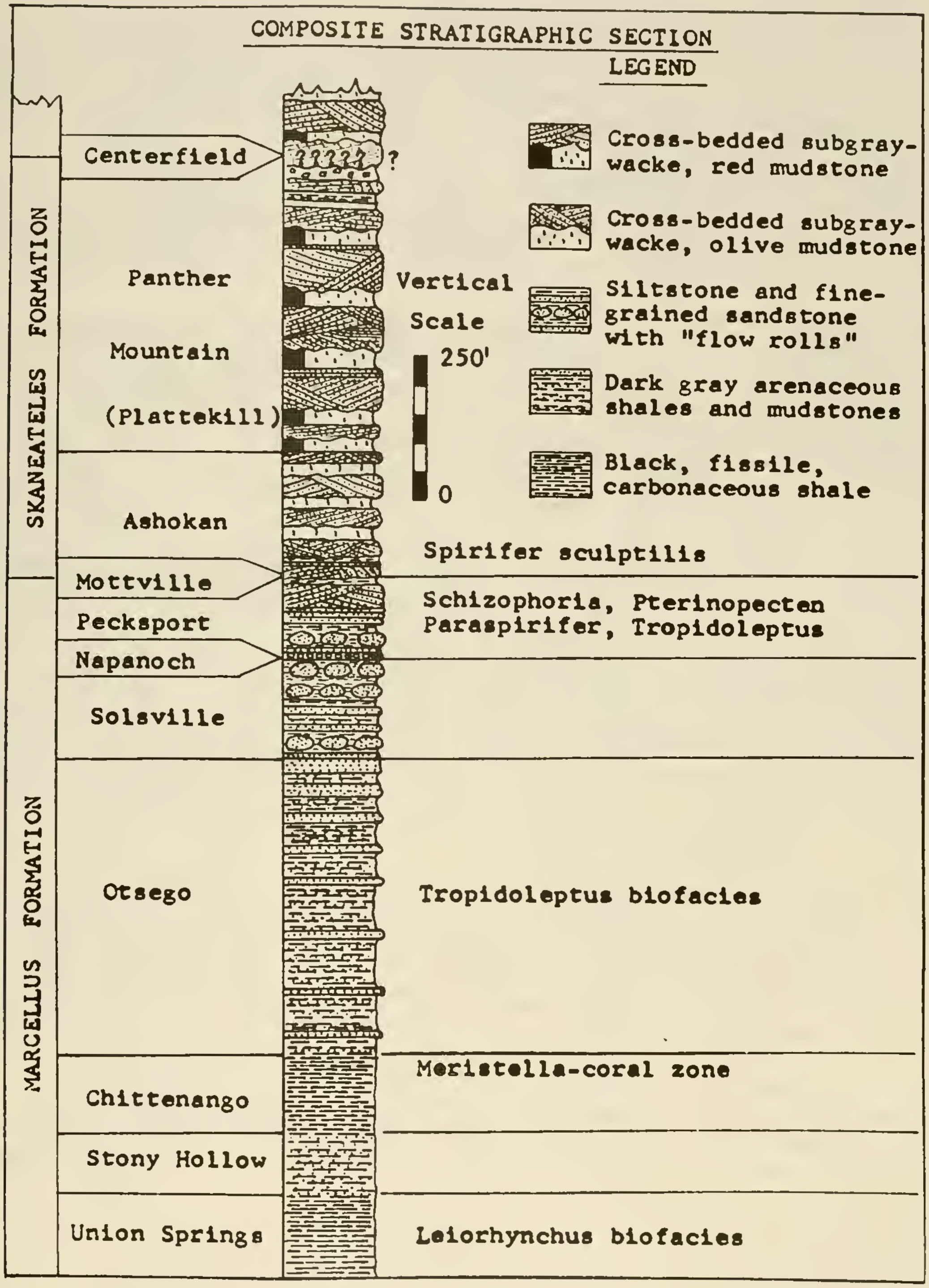


Figure 7. Composite vertical section of Marcellus and Skaneateles Formations in southeastern New York

DELTAIC SEDIMENTATION WITHIN THE MARCELLUS FORMATION

The Vertical Profile

"Each fundamental sedimentation process produces a specific environmental distribution of sediment and a specific vertical profile which can be used as a standard in interpreting the geologic section" (Visher, 1965). The "standard" vertical profile for the Marcellus and Skaneateles Formations is illustrated in Figure 7.

In New York and Pennsylvania, the Onondaga-Union Springs contact is a surface of disconformity and erosion (Chadwick, 1927, Willard, 1936). The erosion is slight and may only be local, but it does indicate shallow seas were initially present.

The deposition of the Union Springs, Stony Hollow, Chittenango, and Otsego Members can be described as a typical "fill in" (Krynine, 1959) marine sequence or an upward coarsening prograding deltaic deposit containing the physical attributes of the Cleveland, Moscow, and Big Bend phases of deposition. During this interval sedimentation was greater than subsidence and the seas remained relatively shallow during the formation of these deposits.

The change to features characteristic of the Solsville Member (Smethport phase) also indicates the change from basin and slope to shelf deposits and a change from a predominance of shale to a predominance of massive and cross-bedded sandstones. Zones of "ball and pillow" structures are also characteristic, and have also been noted in the Upper Devonian in many parts of the state. The persistence and fabric of the marine conglomerate in this zone supports the contention of Soruaf (1965) that these are channel deposits which in this case have been reworked and redistributed in the surf zone.

The Napanoch Conglomerate consists of poorly sorted but moderate to well-rounded sub-spherical white quartz and quartzite pebbles and oblate and prolate spheroidal and bladed dark gray and green siltstone and black chert pebbles. The quartz and chert pebbles have a maximum diameter of about an inch, but are usually 0.5 inches or smaller. The siltstone pebbles have a maximum size of 4.5 inches in their longest dimension and frequently lie parallel to the strandline. The entire horizon is no more than eight inches thick, with the pebbles dispersed through reworking in a matrix of coarse and medium sand, and some broken shell fragments. This unit marks the end of the upward coarsening Solsville sequence, and is in close proximity to the shoreline.

The overlying strata, now within the Pecksport Member, indicate the progressive change from littoral to subaerial conditions. Massive fine-grained sandstones and "ball and pillow" structure horizons grade upward into laminated and thin-scale planar cross-bedded subgraywackes containing alpha, beta, and gamma cross-stratification (Allen, 1963). Interbedded dark gray shales and mudstones, some associated with flaser bedding, are also common. The sediments are characteristic of the Smethport phase and indicate lateral as well as vertical migration of several broad tidal channels on the delta platform. The appearance of dark shales and olive mudstones indicates the association of lagoons and tidal flats with the channel deposits.

The only evidence of a transgression is the relatively thin reworked calcareous subgraywacke in the shape of an offshore bar, (Mottville Sandstone). The orientation of the bar, the steeper inclination toward the landward side, and the associated lithologies and structures (Figure 8) all indicate an orientation parallel to the depositional strike. Successive bars may exist seaward and produce a net offlap in the sequence. This interval represents a time of lesser sedimentation when during subsidence, sediments had more of an opportunity to be reworked, transported, and deposited by longshore and tidal currents.

The associated sediments and structures about these bars are characterized by their variability, for no two sections are similar - Figure 8 represents an attempt at a section, but is based on a composite from several sections. The variable lithology at the base of the bars and in the lateral gradations, and the variable position with respect to the last marine horizon tentatively suggest that the feature represents an offshore bar in the Schoharie Valley region, a beach and barrier bar complex north of Kingston, and a series of beach and chenier deposits in the Ashokan area; sufficient data concerning the lithosome geometry is not available to substantiate these inferences.

The overlying shift into the fluvial environment of the Catskill phase, recognized as the Plattekill Formation (Fletcher, 1963) is not only time-transgressive but transitional. The lithologies, contact relations and fabric of the associated sediment bodies, disregarding the presence of redbeds, are quite similar and one is hard pressed to find methods of distinguishing members within this formation on criteria other than color.

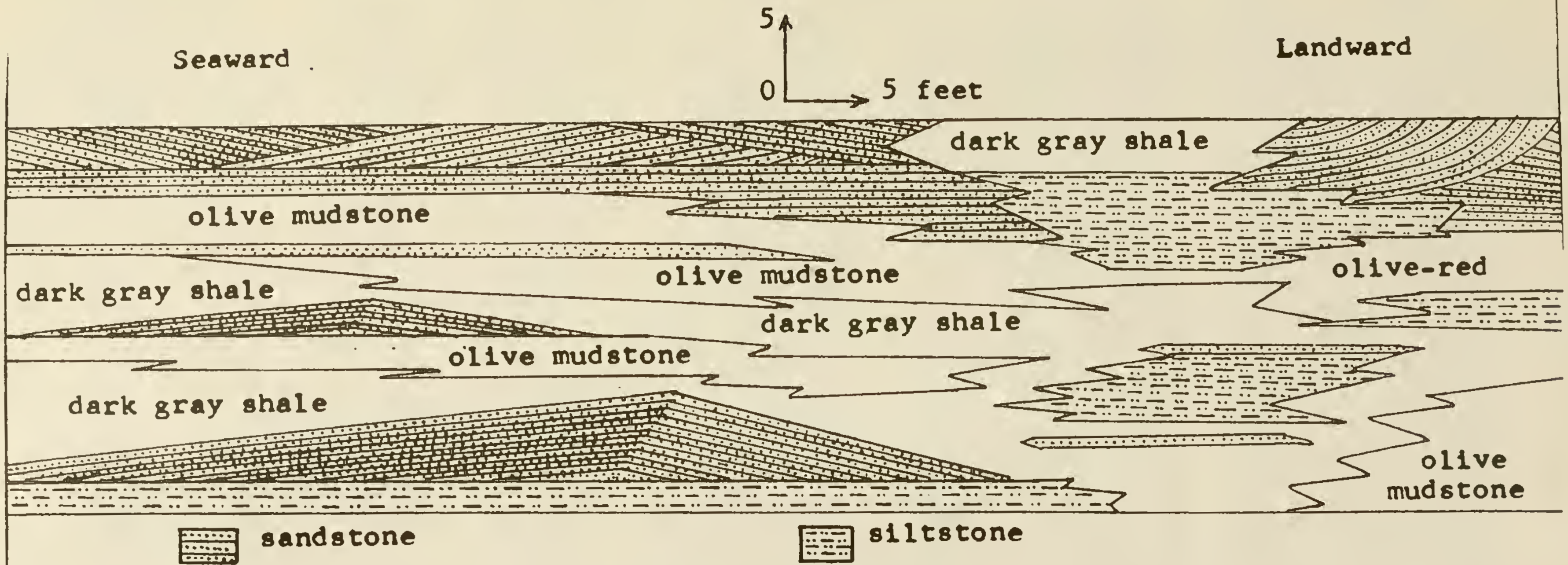
Lateral Sediment Associations

The lateral sediment associations, based on the composite sections (Figure 5) and other outcrops are not presented in any great detail. For more complete information the reader is referred to Wolff (1967a).

The variable thickness of the marine lithofacies between Schoharie Valley and Alcove Reservoir, as well as the northward coarsening of sediments and the presence of dull maroon beds in the vicinity of Towpath and Petersburg Mts. suggest that much of the Panther Mt. Member has had a northern as well as eastern source. The inclination and orientation of the longshore bar at Franklinton also suggests a landmass to the north or northeast. Thus, much of the Panther Mt. may be a prograding delta slope and platform deposit (Smethport phase) building southwestward into the basin.

The Cleveland, Moscow, and Big Bend phases do not exhibit large variations in thickness from Schoharie Valley to the vicinity of

DIAGRAMATIC REPRESENTATION OF ZONE OF BAR ACCUMULATION



ASSOCIATED FEATURES

Coquinites and shale pebble conglomerates Protoquartzites Ripple marks and cross-laminations Megaripples Primary current lineation Plant fragments		Coquinites and shale pebble conglomerates Subgraywackes Rippledrift and mudcracks Tree molds, worm tubes Planar cross-bedding Brackish water fauna
Medium-grained sandstone Fair sorting Gentle dip (2-8°) Dark arenaceous shale	Coarse-grained sandstone Good sorting Steep dip (8-24°) Olive mudstone	Siltstone Poor sorting Horizontal bedding Olive-red mudstone

Figure 8. Local lithofacies associated with Mottville Sandstone (based on information from several localities).

Ashokan Reservoir, and in general, lateral variations are quite indistinct. It is only from the zone of "ball and pillow" structures upward (largely in the Smethport phase) that lateral variations become dominant.

The area in the vicinity of Alcove Spillway contains an anomalously thick section (Figure 5), a marine re-entrant well above the Mottville Sandstone, and lack of "redbeds" for another 250 feet in the section. This embayment (Albany Bay of Goldring, 1943) appears to be tectonically controlled for marine conditions persist above coarse subgraywacke and interbedded shales (Figure 9).

The area east of Ashokan Reservoir contains a Marcellus marine fauna in dark arenaceous shales above the initial sequence of cross-bedded subgraywackes, but below the Mottville Sandstone. The presence of Schizophoria in these lagoonal deposits (Cooper, in Goldring, 1935) led to the recognition of the Pecksport Member in this region and the appearance of marine conditions above the first channel sandstones again suggests a period during which subsidence dominated over sedimentation.

Several other features indicate a rapid influx of sediments from the northern margin of a large clastic wedge, centered in the vicinity of Ellenville, that spread into the basin. These include:

a. An increase in the thickness of the Stony Hollow, Otsego, and Solsville Members as well as an increase in the Smethport phase (initial redbeds do not appear until one is 320 feet above the Mottville interval).

b. The progressive southward increase in the size and abundance of pebbles and pebble horizons, and flow rolls, and flow roll horizons.

c. The presence of a shallow tidal channel trending southeast truncated by the subgraywacke Mottville Sandstone and a series of blanket sand deposits (cheniers) interbedded with olive mudstones and shale-pebble conglomerates.

These features suggest that tectonic activity may have increased to the southwest. A generalized interpretation of the lithofacies and paleogeography of the Mottville interval across this region is shown in Figure 9. The field trip includes a vertical section of the Marcellus sequence and will attempt to cover several of the lateral sections associated with the Mottville Sandstone.

Acknowledgements

The writer would like to thank Peter Buttner for accompanying him in the field and is appreciative to Dr. L.V. Rickard and Julian Kane for reviewing the content of this report.

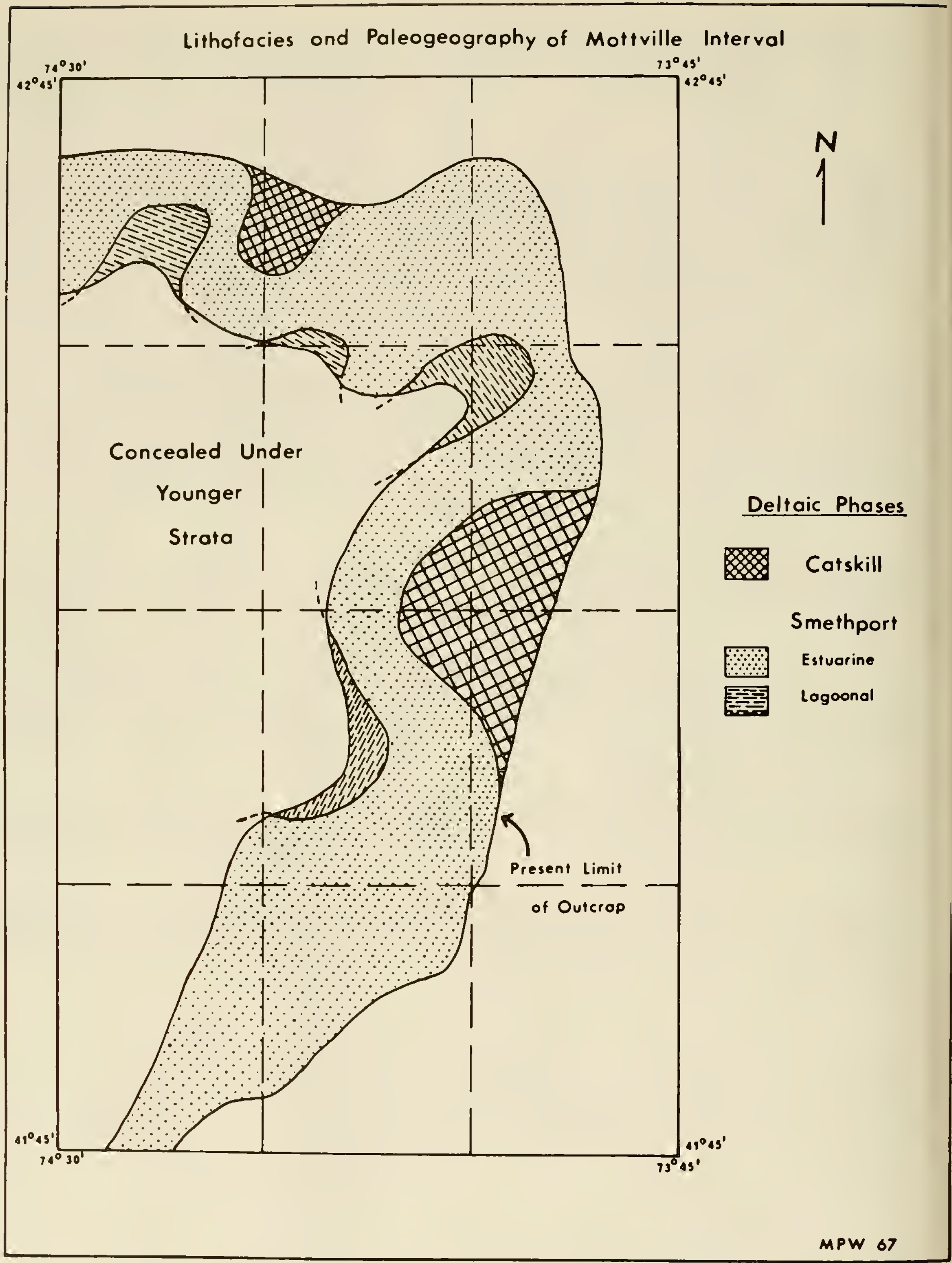


Figure 9. Location of embayments in Smethport phase during deposition of Mottville Sandstone

Devonian Marcellus Fm. Trip - Road Log and Stop Descriptions

We will drive the maximum distance to be encountered at any single time to Stop 1. (nearly 60 miles on the Thruway-about one hour). The remainder of the day will bring us closer to the point of origin. The general route and location of stops are noted in Figure 10 on the next page.

MILEAGE

Total:	Distance	
0.0	0.0	Turn right (north) onto Washington Ave. in front of the Thruway Motor Inn and follow signs to N.Y. Thruway.
1.2	1.2	Turn sharp right onto ramp that leads into Thruway.
1.7	0.5	Follow signs overhead that lead to Thruway-South (N.Y.C.).
	0.5	Toll booth at Exit 24; pick up card and follow left fork to New York City and South. Toll to Kingston (Exit 19) is 90¢.
6.6	4.4	Pass by Albany Exit 23, continue straight on Thruway.
34.8	28.2	Pass by Catskill Exit 21; you are now half-way.
55.0	20.2	Exposure of Chittenango Member on west side of Thruway.
	2.0	Outcrop of Stony Hollow Member on west side of Thruway.
57.8	0.8	Toll booth at Kingston Exit 19 (90¢ from Exit 24).
	0.1	Go ¼ way around traffic circle and follow overhead signs to Route 28-north (avoid going on Rt. 209).
58.8	0.9	Stop 1. Cliff face at intersection of Rt. 28 with road to top of Sky Top Motel. Turn right and park along road at base of cliff.

Stratigraphic units: Union Springs and Stony Hollow.
Deltaic phases: Cleveland and Moscow

Description: Only the upper 20 feet of the calcareous, black, fissile shales of the Union Springs are exposed. Though fossils are rare, thin limestone layers or black calcareous concretions (2-3 feet in diameter) are sometimes noted in the upper portion of the unit. Westward overthrusting and shearing have produced quartz-filled tension fractures and local zones of slickensided vitreous, black shales, especially near the contact with the Stony Hollow.

The contact is a broadly undulating surface overlain by somewhat more resistant strata. Only 35 feet of the interbedded calcareous shales are exposed - note how the lithology controls the changes in fracture pattern and surface weathering between the two members.

Interpretation: The Union Springs, though rarely exposed in this region, shows little vertical or lateral variation and exhibits features characteristic of the Cleveland depositional phase. Its proximity to the underlying Onondaga Limestone indicates it is a euxinic shallow-water deposit. The Stony Hollow Member can be more clearly observed and interpreted at Stop 2.

59.1	0.3	Continue northwest on Rt. 28; stop at the steep cliff on the north side of the road.

Figure 10. Route and location of road stops and driving time between Albany and first and last stops.

Stop 2. Type section of Stony Hollow Member
Stratigraphic unit: Stony Hollow
Deltaic phase: Moscow (Chagrin in part?)

Description: This section consists of 130 feet of calcareous siltstone and shale. The silty units dominate the sequence, are more calcareous, and weather to a buff-orange (the carbonate may be of detrital origin). Individual beds vary from one inch to more than one foot in thickness. Weathering accentuates the vertical lithologic contrasts of interbedded siltstone and shale and points out the persistence of the irregular laminations. These features could suggest deposition from density currents, but this is not clearly supported by the internal structures. The vertical contacts between the cyclical depositional units are not smooth and sharp but irregular and gradational (no strong erosional contact). The basal unit is not graded, though the overlying portion may show irregular and micro-crosslaminations, and is, in turn, overlain by a thin shale seam. The micro-crosslaminations are not continuous nor do they maintain a persistent orientation. Much of the silt-sized material is carbonate and may be of detrital origin.

Interpretation: Most of the sequence probably formed by deposition from normal marine currents, but the presence and persistence of "subdued" turbidite features could also suggest that portions of the unit may have been redeposited as slow moving density currents. The abundance of carbonate material places this member in the Moscow phase of sedimentation.

	0.3	Continue northwest on Rt. 28; exposure of Chittenango Member on south side of road
60.0	0.3	Diner and gas station - Rest Stop (10 minutes)
	0.5	Outcrop of Otsego Member on north and south side of Route 28.
61.2	0.7	Junction with Route 28-A; bear left onto Route 28-A.
	0.3	Exposure of Lower Solsville "ball and pillow" structures (flow rolls) on north side of road.
	0.7	St. John's R.C. Church - Napanoch Conglomerate exposed 100 yards east of church on north side of Rt. 28-A.
	0.2	Junction with Morgan Hill Road (Oehler's Mt. Lodge)- continue straight on Rt. 28-A but note complex planar crossbedded sandstones of Pecksport Member on south side of road.
62.6	0.2	Stop 3-A Junction of Rt. 28-A and road to West Hurley R.R. Station and Rt. 28.

Stratigraphic units: Pecksport and Mottville
Deltaic phase: Smethport

Description: We have passed through the prodelta slope and distal delta platform deposits in this region to concentrate on the neretic, supratidal and alluvial environments of the proximal delta platform. Note the numerous lateral and vertical changes in contact relations, lithology, and sedimentary structures across the outcrop.

The base of the exposure consists of flat-bedded subgraywacke flagstones on the north (right) while arenaceous dark gray shales occur at the southern end. This is overlain by a discontinuous, lens-shaped dark shale truncated by a trough-shaped erosion surface. The trough is filled with thick sets of planar cross-bedded that grade into nearly horizontal interbedded sandstones and shales on the south. The lowest point in the trough contains coarse sands with numerous shale clasts and plant fragments and indicates a local (anomalous) relief of 6-7 feet,(Figure 11).

The sequence is truncated by a slightly undulatory contact and is overlain by a moderately-sorted calcareous subgraywacke with long planar crossbeds dipping to the northwest. The unit becomes coarser vertically and decreases in thickness to the north. The upper portion of this unit contains smaller sets of wedge-shaped planar crossbeds separating local scour and fill surfaces.

A few inches of gray shale overlie this unit and is capped by a thin, irregularly laminated subgraywacke, 1-2 feet thick, that grades upward into a gray-green knobby silty mudstone. This is followed by another laminated, moderately-sorted subgraywacke - knobby green mudstone sequence.

Interpretation: The planar crossbedded sandstones across the erosion surface and the vertical and lateral changes in texture and structure indicate a tidal or supratidal channel, sloping to the southeast, and undergoing lateral migration from point bar deposition toward the south (Figure 11). No marine fossils have been found at this outcrop.

The contact with the moderately-sorted Mottville Sandstone marks the beginning of an interval of sediment reworking and the formation of thin beach or barrier bar deposit. The slope of the channel indicates transport to the southeast - exactly <u>opposite</u> to the known land-sea relations in this region. This is believed to be a local feature related to the folding and uplift that occurs in slightly younger strata at the Ashokan Reservoir Spillway southwest of this exposure, and suggests that some tectonic adjustments were contemporaneous with deposition during the formation of these units. All other regional and local dispersal features indicate transport to the west and northwest.

The thin, laminated subgraywacke above the barrier bars are also persistent, and their gradation and association with the green-gray silty mudstones (upward-fining sequence) and numerous plant fragments suggests they may be cheniers.

62.7 0.1 <u>Stop 3-B</u> Continue north on the West Hurley Road toward the junction with Route 28.
Stratigraphic units: Mottville and Ashokan
Deltaic phases: Smethport and Catskill

Description: The lower exposure is the bar-like continuation of the Mottville Sandstone. The basal contact with gray arenaceous

Figure 11. Channel with slope opposite regional dispersal patterns beneath Mottville Sandstone on Route 28-A (Stop 3-A)

Figure 12. Interbedded wedge-shaped sandstone and shales in Ashokan railroad cut (Stop 4)

shales is sharp while the upper contact shows the superposition of large and small-scale ripples and another sharp contact with the overlying orange-weathering, knobby, olive mudstone.

The upper exposure is composed of two laminated, moderately-sorted subgraywackes, three to four feet thick, with ten feet of a dull maroon - olive silty mudstone between them. Note the gradational contacts between the two units.

Interpretation: These sandstones are the landward extension of subgraywackes on Route 28-A and again are enclosed in nonmarine strata. The local effects of uplift are reflected in the oxidation of the silty mudstones - these are the only red sediments for another 350 feet in the section. The upper units are considered cheniers because of their sorting, laminations and association with olive and red mudstones. This section is overlain by scoured surfaces and with planar and trough crossbedded subgraywacke channel sandstones dispersing sediments northwestward into the basin.

0.3 Continue north on road to Route 28; note gray channel sandstones of Ashokan Member on south side of road.

63.5 0.5 Stop 4. Turn right onto deadend dirt road just before railroad overpass and junction with Route 28.
Stratigraphic unit: Ashokan Member
Deltaic phase: Catskill

Description: Section consists of several "upward fining" cycles characteristic of channel fill sediments in alluvial sequences. Unlike the typical Catskill phase, there are thick sequences of planar crossbedding, numerous interbedded siltstones and shales, a lack of redbeds, and the general absence of a clear vertical zonation of plane beds, large and small scale ripples, and other sedimentary structures that characterize alluvial sequences.

The channels lie on erosion surfaces of asymmetric ripples and are filled with planar crossbedded graywacke and subgraywacke also containing horizons of shale intraclasts, concretions, and abundant plant fragments. Many of the interbedded gray siltstones and dark arenaceous shales are lens-shaped and discontinuous due to the lateral erosion surfaces. The interbedding between sandstone and shale units is quite prominent (Figure 12).

Interpretation: The planar crossbedding of subgraywackes and the wedge-shaped interpenetration of fine-grained sandstones and shales reflect the lateral migration of point-bar deposits. This is also evident by the lateral erosion surfaces across discontinuous shale sequences. The presence of vertical and lateral accretionary deposits suggests the formation and migration of alluvial channels across a supratidal sequence.

0.1 Exit and continue north beneath railroad overpass to junction with Route 28. Turn right - you are now heading east back toward the Thruway.

0.5 Outcrops of Ashokan units previously seen in R.R. cut.

64.3 0.2 Stop 5. In parking area at base of road cut on south side of Route 28.
Stratigraphic units: Pecksport - Mottville - Ashokan
Deltaic phases: Smethport and Catskill

Description: Base of section contains 22 feet of irregularly stratified and cross-laminated brownish siltstone and interbedded black shale. Ripple marks, shale clasts, worm tubes, and a marine fauna with characteristic Pecksport fossils are also present. The shales are overlain by a planar cross-bedded and laminated moderately sorted subgraywacke, about seven feet thick. The upper portion contains some scour and fill surfaces containing wedge-shaped sets of planar cross-beds with some change in orientation or inclination. The contact with the overlying shale is sharp and a sequence of interbedded massive sandstones and shales, 3-6 feet thick follows. These are sharply truncated by a series of northwest trending erosional channels filled with trough cross-bedded sandstones and siltstones that connect with the section exposed in the railroad cut.

Interpretation: This outcrop lies one mile northwest of Stop 3 and is more typical of the Mottville exposures north and south of the Ashokan area. The persistence of marine conditions above the first cross-bedded sandstones (35 feet below this section) and behind and below the barrier bar indicate the development of a lagoon or embayment. The widespread and steep erosional channels above this sequence indicate the rapid change to prograding intertidal and alluvial conditions that persist well into the Upper Devonian of the Catskill front.

	0.6	Continue east on Rt. 28; base of Pecksport, with planar-cross-bedded sandstones is on north side of road.
	0.2	Exposure of upper Solsville Member with Napanoch Conglomerate on north side of road.
	0.3	"Ball and pillow" structures at base of Solsville.
65.7	0.3	Junction of Rt. 28 with Rt. 28-A. Travel "loop" is now complete; continue east on Rt. 28 to Thruway entrance.
	3.1	Toll booth at Thruway. Go north to Saugerties - Rt. 32 (Exit 20)
80.0	10.2	Exit 20 on Thruway (20¢ toll); turn left and follow signs to Rt. 32 - north (Hunter & Palenville).
	0.2	Turn right onto Route 32-north.

83.0 2.8 Stop 6-A. Park on right (north) side of Rt. 32 about 1/3 of way up the Hoogeberg ridge.
Stratigraphic unit: Otsego
Deltaic phase: Big Bend

Description: Lower outcrop (6-A) consists of 1-6' sequences of interbedded blocky arenaceous shales and massive siltstones. Laminations and other internal sedimentary structures are not common and most contacts are gradational. The abundance and thickness of the sandstones increases vertically in the section.

83.2 0.2 Stop 6-B. Farther along Hoogeberg escarpment on Route 32.
Stratigraphic units: Solsville-Napanoch-Pecksport
Deltaic phases: Big Bend and Smethport

Description: Upper outcrop (6-B) contains a contorted, massive, argillaceous sandstone, about eight feet thick, as the basal unit. Most of the 75 feet of the section is composed of these massive units, usually 2-3 feet thick, deformed into "ball and pillow" structures with the dark interbedded shale squeezed and deformed about them. Several coquinite horizons are present, as well as a thin but laterally persistent marine conglomerate. The deformed structures as well as the undeformed massive sandstones continue to dominate the higher portions of the sequence. These may weather a dull maroon and some contain red and black shale fragments or small concretions near their upper contact. Mucrospirifer and Tentaculites persist to the top of the section.

Interpretation: (6-A and B) The upward coarsening progradation is evident in the lower (Otsego) as well as the upper (Solsville) exposures. The persistence of these upward coarsening units, as well as the conglomerate and "ball and pillow" structures along the depositional strike for 80 miles indicates the wide extent of this vertical zonation, and is used to define the prodelta slope and marine platform of this deltaic sequence. No attempt has been made to correlate tectonic events through the tracing of "ball and pillow" structure horizons but this seems possible if more sections were available.

83.5 0.3 Stop 6-C. Near crest of Hoogeberg escarpment on Rt. 32.
Stratigraphic units: Mottville and Ashokan
Deltaic phases: Smethport and Catskill

Description: The top of the marine section with the massive laminated fine-grained sandstones and "ball and pillow" structures is overlain by 14 feet dark gray barren shales and medium simple-crossbedded and flatbedded subgraywacke (most have been quarried as "flagstone") with abundant plant fragments.

This is overlain by massive gray siltstone and laminated, calcareous, planar cross-bedded , well sorted subgraywacke, about five feet thick, lying on a wave scoured erosion surface and containing a basal intraformational conglomerate of shale clasts. More complex scour and fill and planar cross-bedding occur near the top of the unit. It is overlain by two feet of olive-gray knobby mudstone and a sheet-like laminated moderately sorted subgraywacke,about one foot thick. A steep erosional contact follows, above which are laminated planar crossbedded sandstone also containing the mold of a portion of the Middle Devonian tree-fern, Eospermatoperis (Figure 13). These sands are also truncated by erosional surfaces higher in the section (Figure 14).

Figure 13. Eospermatoperis mold in channel sandstone above Mottville Sandstone on Rt. 32 (Stop 6-C)

Figure 14. Truncated planar cross-beds above Mottville Sandstone on Route 32 (Stop 6-C)

Interpretation: The description (and the interpretation) is quite similar to Stop 5 on Route 28. The Mottville Sandstone is again truncated by channel deposits also containing tree remains, and these, in turn are overlain by other channel deposits - all sloping to the northwest.

This is the last outcrop before the lunch stop (distance of 10 miles). The next exposure to be visited is on the other side of the basin - 30 miles from this section.

93.1	9.6	Lunch Stop - Mike's Diner on Route 32.
96.3	3.2	Alternate Lunch Stop - Rockface Diner on Route 32.
96.5	0.2	Intersection of Route 32 with Routes 23 and 145 (traffic light). Turn left (west) and park when ready to continue field trip. We will meet here at a time designated by field trip leader.
	1.5	Exposure of channels and red flood plain deposits of Plattekill Member.
98.2	0.2	Traffic light - turn right (north) on Route 32. Routes 23 and 145 continue straight ahead.
	1.0	Steel bridge on Route 32 across Plattekill Member
102.7	3.5	Town of Freehold
107.0	4.3	Junction with Route 81 in Greenville-continue straight on Route 32.
113.4	6.4	Junction with Route 143 - turn left onto Route 143.

113.6 0.2 Stop 7-A. Park at Westerlo firehouse on Route 143.
Stratigraphic units: Solsville-Napanich-Pecksport
Deltaic phase: Big Bend

Description and Interpretation - see 7-B

113.9 0.3 Stop 7-B. In quarry on top of hill on left (south) side of Route 143.
Stratigraphic unit: Pecksport
Deltaic phase: Big Bend

Description: (7-A and B) A sequence of "ball and pillow" structures with the polymict marine conglomerate in interbedded fine-grained sandstones and shales occurs 0.2 miles down the road. This section at the firehouse, normally composed of laminated fine-grained sandstone and "flagstones", is similar to the one below, and even contains another conglomerate. Marine arenaceous shales and massive laminated siltstones continue to persist above this outcrop and the section (and ridge) is capped by a laminated sandstone interbedded with shales containing several coquinite horizons that suggested the Pecksport Member to Cooper (in Goldring. 1943).

Interpretation: The persistence of marine conditions for over 100 feet above the conglomerate is unknown elsewhere in the Marcellus Formation. The splitting of the conglomerate into two units, the presence of massive laminated marine sandstones, and the persistence of several coquinites is also unknown in the sections along the Hoogeberg escarpment.

The persistence of marine conditions through a greater vertical section suggests the presence of a tectonically controlled embayment (Albany Bay of Goldring, 1943). The splitting of the conglomerate suggests a change and addition of another source area, and, based on the stratigraphic position of "flagstones" in the Reidsville quarry north of this stop, the development of another delta lobe from the northeast. Thus, the Pecksport Member is abnormally thick in this region due to the greater sedimentation that accompanied subsidence. Yet, near the top of this unit, and in the quarries at similar elevations throughout the Berne-Westerlo region, the abundance and persistence of coquinites and general lack of "ball and pillow" structures indicates that there were also periods of great stability.

In summary, this region is a tectonically controlled embayment that occurs in the transition between the tectonically active sedimentation patterns to the southwest, along the Hoogeberg escarpment, which led to the development of persistent linear features, and the relatively stable deposits striking westward toward Schoharie Valley and central New York which show irregular curving depositional trends.

114.3 0.4 Stop 8. On Route 143 in depression at junction with private road on right.
Stratigraphic unit: Mottville
Deltaic phase: Smethport

Description: Small but significant exposure of the calcareous marine subgraywacke (without *Paraspirifer acuminatus*). It is a planar and flat-bedded sandstone, about five feet thick. The underlying and overlying contacts were not observed.

Interpretation: Its position above the laminated sandstones and coquinites of the Pecksport indicates it is probably a remnant of the Mottville Sandstone.

2.5 Junction of Route 143 with Rensselaerville Road in Westerlo - continue straight, do not stay on Route 143 which turns right.
2.4 Junction with Route 85 - turn left onto Route 85.
4.6 End of Route 85 in Rensselaerville; turn right toward Route 145 and Livingstonville on County Road 353.

124.0 0.2 Stop 9. Edmund Niles Huycks Preserve - park at white house before bridge over Rensselaerville Falls. (NO HAMMERS AT THIS STOP - we will spend one hour climbing the falls and returning by a woodland trail to the entrance.
Stratigraphic units: Pecksport-Mottville-Panther Mt.-Plattekill
Deltaic phases: Big Bend, Smethport, Catskill

Description: Section below the bridge contains 60 feet of massive argillaceous sandstones and shales with a few Marcellus fossils. The section above the bridge consists of 30 feet of interbedded dark shales and siltstones with a fauna suggestive of the Pecksport.

Straight and linguoid ripples occur with their long axes trending N.50°W. The first falls are capped by an en-echelon series of fine-grained bar-shaped sandstones 3-4 feet thick, whose crests run parallel to the ripples. The cross-beds dip 3-4° toward the seaward side (up the falls) and about 17° toward the strandline (downstream). Both contacts are sharp; the barrier bars are overlain by 50 feet of medium-gray arenaceous shales and siltstones (no marine fauna noted). In ascending the falls, note the gradual color change into 35 feet of dull maroon, pale green, and finally red. These shales are truncated by an erosion surface that is overlain by eight feet of dull red subgraywacke with basal red and green shale fragments and limestone clasts.

Interpretation: As at Dormansville, but unlike the sections along the Catskill front, there is a greater transition interval preserved between the last marine strata and the first channel sandstones. The small size and finer texture of these bars indicate a more distal position from the source area. The persistence of shales and siltstones above these units, and the gradual change from gray to red could indicate an environmental shift from proximal marine to barrier bar to interdistributary bay (another embayment) to alluvial floodplain. Similar features seem to persist through the Panther Mt. Member in the Schoharie Valley region.

	0.2	Upon leaving the Preserve, cross the bridge and continue uphill on Albany County Road 353.
	0.3	Take right fork near top of hill (County Road 353) to Livingstonville. Most exposures along this road are within the Plattekill Member.
	3.3	Junction with road to Crystal Lake - continue straight on County Road 353.
132.4	4.6	Junction of Rt. 353 with Route 145 in Livingstonville; turn right onto Rt. 145.

136.2 3.8 Stop 10. Southern end of Mottville Sandstone on east side of Rt. 145 (outcrop continues for 0.3 miles).
Stratigraphic units: Mottville and Panther Mt.
Deltaic phase: Smethport

Description: Exposure consists of 18 feet of medium-sized, planar cross-bedded and laminated calcareous subgraywacke. Portions of the unit become coarser vertically and contain several small wedge-shaped erosion surfaces with planar cross-beds. Most of the beds dip to the north and northeast (toward shore) and may be cut by these scour and fill surfaces. The middle of the section is overlain by dark marine shales with a Marcellus fauna, but *Spirifer sculptilis* has not been found.

Interpretation: This is the largest and most extensive exposure of the Mottville Sandstone, and its shape, texture, and internal structures suggest a reworking of nearshore sediments during the formation of the Mottville Limestone, and the development of a submerged barrier bar or offshore bar or beach near another marginal embayment.

140.9	4.7	This was the last stop; the return to the Thruway Motor Inn in Albany will take slightly more than one hour. Continue north on Rt. 145 toward Middleburg.
	1.2	Junction of Rt. 145 with Rt. 30 in Middleburg. Leave Rt. 145 and turn right (north) onto Rt. 30 toward Schoharie.
148.3	6.2	Junction of Rt. 30 with Rt. 43; turn right (east) onto Route 43.
	3.9	Junction of Rt. 43 with Rt. 146; turn left (north) onto Route 146.
	11.4	Intersection of Rt. 146 with Western Ave. in Altamont-turn right.
	0.4	Junction with Rt. 146 and Main Street; turn left onto Main St. (Rt. 146).
	2.5	Intersection of Rt. 146 with Rt. 158 - continue on Rt. 146 through Guilderland Center.
170.0	3.5	Junction of Rt. 146 with Route 20; turn right (east) onto Rt. 20 towards Albany.
	4.5	Junction with road to Rt. 87 (Thruway) - turn left.
	0.8	Take right fork at sign to "Office Buildings".
175.7	0.4	Get off of Thruway spur road at Washington Ave. (Exit 2; turn left at light.
176.5	0.8	Entrance to Thruway Motor Inn on Washington Ave.

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Trip 21

UNCONFORMITIES AT THE NORTHERN END OF THE BERKSHIRE HIGHLANDS[1]

by

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Quadrangle required: Windsor, Massachusetts, 1960, $7\frac{1}{2}$ minute.

NOTE: Stops 4 and 5 will require moderately strenuous traverses.

INTRODUCTION

The Windsor quadrangle straddles the axis of the Berkshire Highlands of Massachusetts. Cambrian and Ordovician miogeosynclinal clastic and carbonate rocks (henceforth called the "western sequence") are exposed in the western third of the quadrangle. The eastern third of the quadrangle is underlain by Cambrian and Ordovician eugeosynclinal volcanic rocks, pelites, and graywackes. All of the Paleozoic rocks have mineral assemblages consistent with the garnet or kyanite zone (Norton, 1967). The central part of the quadrangle is underlain by feldspathic gneisses and minor amounts of calc-silicate gneiss, amphibole gneiss, graphitic gneiss, and quartzite; all of probable Precambrian age. These Precambrian rocks were subjected to at least sillimanite (+muscovite) grade metamorphism. Radiometric ages on correlative rocks in Vermont date this event as being equivalent to the late Precambrian Grenville orogeny (0.9 b.y.; Faul and others, 1963; Lyons and Faul, 1968). Relict evidence for this earlier metamorphism has been partially obscured by mid-Paleozoic (Acadian) metamorphism which retrograded the Precambrian mineral assemblages. The metamorphic discontinuity between the Precambrian and Paleozoic rocks is best seen in the calc-silicate gneiss.

[1] Publication authorized by the Director, U. S. Geological Survey

The dominant Paleozoic structure of the quadrangle is the Hoosac nappe. This Acadian recumbent anticline is overturned to the west and has three large amplitude satellitic syncline on its inverted limb. The term Hoosac nappe is restricted to those rocks to the east of the Savoy Hollow Brook thrust (Figure 1) although structures to the west of the same thrust suggest that nearly the entire quadrangle is involved in the nappe. In the northwestern part of the quadrangle, the western sequence of rocks is inverted in the overturned limb; in the southwestern part of the quadrangle, the western sequence is upright in what may be the normal limb of a major syncline overturned to the west. This fold is probably related to the Hoosac nappe. Syn- and perhaps post-folding thrust faults are major features which have complicated the structure of the nappe.

The Windsor quadrangle is three miles south of the northernmost exposures of the Precambrian core of the Berkshire Highlands and thus is one of the best areas to study the stratigraphic relationships between the eastern and western rock sequences. Furthermore, the structural relationships in the quadrangle bear strongly on the problem of a source area or root zone for the large mass of Taconic allochthonous rocks that make up Mount Greylock to the west. If Zen (1967) is correct in proposing that the Taconic rocks were originally deposited on the site of the presently exposed Precambrian rocks, a problem arises in the Windsor area and adjacent areas in that the proposed source area is too small to yield the quantity of known Taconic rocks. Furthermore, miogeosynclinal facies rocks (in particular the Cheshire Quartzite; see Figure 2) are now within 2 miles of eugeosynclinal rocks in the eastern extension of the Savoy Hollow Brook syncline and the Taconic rocks to the west of the quadrangle that would have been derived from this area are more than 10 miles wide. Clearly, there is not sufficient room from which to derive the Taconic rocks in this area. The answer to this dilemma must lie in extensive east-west shortening of the crust, which must have been accomplished in large part by overfolding and intraformational east-over-west shear and in part by thrusting.

The purpose of Trip 21 is to:

1. Look briefly at the rocks exposed in the area, primarily from a stratigraphic viewpoint.
2. Examine the Precambrian-Cambrian contact and the basal rocks of the Paleozoic section. This is relevant to the problem of east-west correlation and also has bearing on the source area for the Lower Paleozoic clastic rocks.
3. Examine the relationships between the Berkshire Schist of Middle Ordovician age (Figure 2) and the various units upon which it rests unconformably. The Berkshire is a clastic unit deposited during the time that the area now underlain by the Precambrian rocks was apparently emergent. Just prior to the emplacement of the Taconic Mountains, extensive erosion of the western sequence occurred in the vicinity of the Windsor quadrangle; then followed an onlap of the Berkshire facies onto the erosion surface. This relationship is clearly shown in the Windsor quadrangle.

ACKNOWLEDGMENTS

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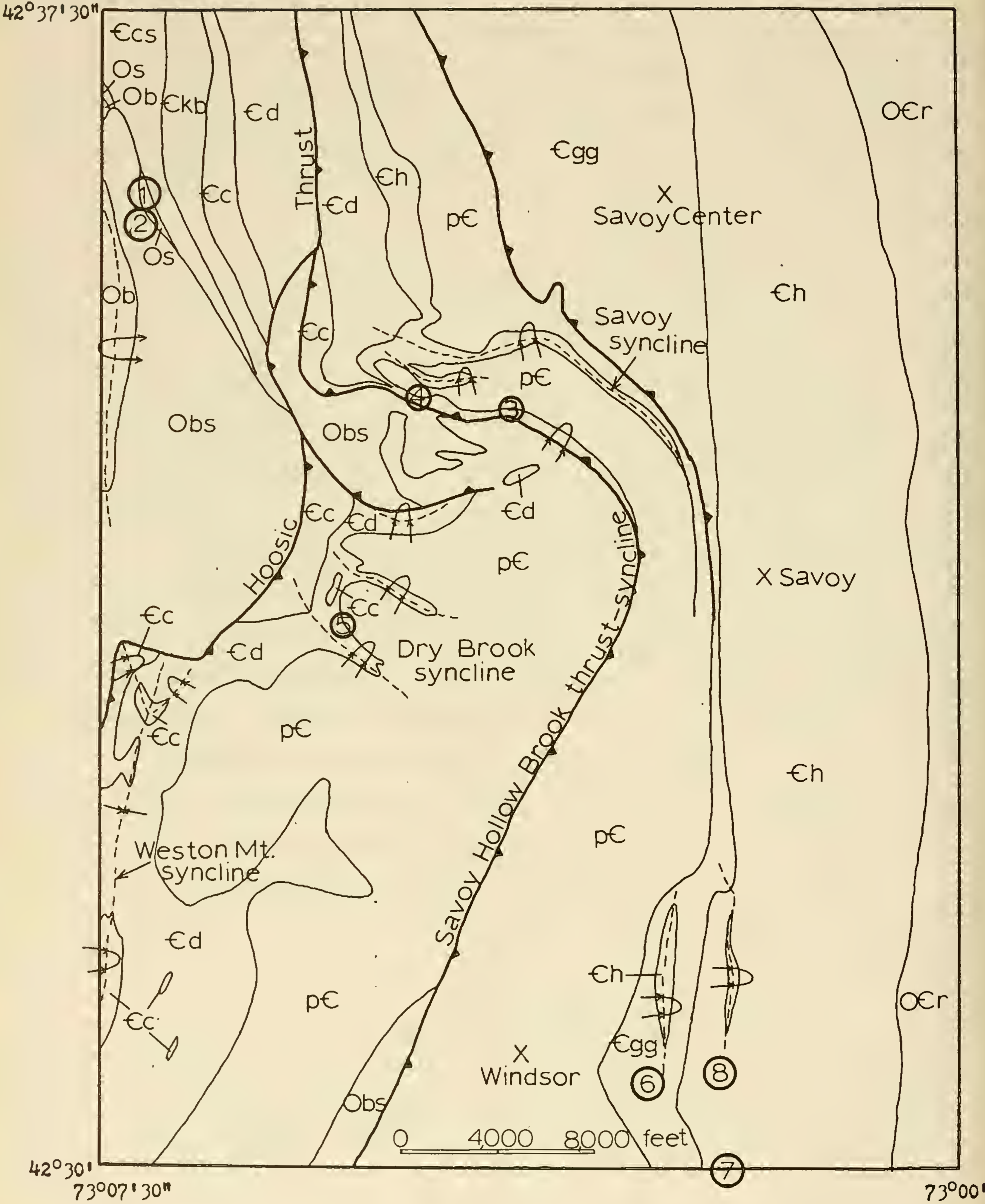

Figure 1. Generalized geologic map of the Windsor quadrangle, Massachusetts.

Explanation of symbols on facing page.

Western Sequence			Eastern Sequence	
Formation	Map Symbol	Age	Formation	Map Symbol
Berkshire Schist	Obs	Middle Ordovician	Not exposed in the Windsor quadrangle	
Bascom Formation	Ob	Lower Ordovician		
Shelburne Marble	Os			
Clarendon Springs Dolomite	Ꞓcs	Upper Cambrian	Rowe Schist	OꞒr
Kitchen Brcok Dolomite	Ꞓkb	Lower and Middle Cambrian		
Cheshire Quartzite	Ꞓc	Lower Cambrian		
Dalton Formation	Ꞓd			
basal conglomerate	Ꞓdc			
Hoosac Formation	Ꞓh	Lower Cambrian or older	Hoosac Formation	Ꞓh
			Granitic Gneiss	Ꞓgg
		Major Unconformity		
Gneiss	pꞒ	Precambrian	Gneiss	pꞒ

Trace of fold axial surface

Trace of thrust fault, teeth on upper plate

4 Stop on itinerary. Dotted line indicates foot traverse.

Figure 2. Explanation for Figures 1, 3, and 4 and stratigraphic column for the Windsor quadrangle, Massachusetts showing the rock sequences west and east of the Precambrian core of the Berkshire Highlands.

L. M. Hall, N. L. Hatch, Jr., P. H. Osberg, L. R. Page, N. M. Ratcliffe, J. B. Thompson, Jr., and E-an Zen. A. K. Gibbs served as a very capable assistant during 1968.

STRATIGRAPHY

Figure 2 gives the stratigraphic column for the rocks of the Windsor quadrangle. The Kitchen Brook Dolomite, Clarendon Springs Dolomite, Bascom Formation, and Rowe Schist will not be seen on this trip and will not be described here. For descriptions of the western sequence the reader is referred to Herz (1958, 1961) and Norton (1967) and for descriptions of the eastern sequence the reader is referred to Hatch and others (1966, 1967, 1968), Norton (1967), and Osberg and others (in press). A brief description of the units to be seen at Stops 1 to 8 follows:

Precambrian Gneiss

The Precambrian rocks are composed predominantly of equigranular, fine-grained quartz-feldspar(normally both plagioclase and microcline)-biotite gneiss. Accessory minerals may include epidote, garnet, and an opaque mineral. Muscovite is typically absent or very sparse. The rock is typically well-banded, both compositionally and texturally. Amphibole gneiss, calc-silicate gneiss, quartzite, and graphitic gneiss locally may predominate.

Hoosac Formation (western sequence)

In the northwestern part of the quadrangle, the Hoosac Formation of Lower Cambrian (?) age or older forms the base of the Paleozoic section. Two different lithologies have been distinguished. A medium-grained, graphitic quartz-muscovite-albite-garnet-biotite(or chlorite, or chlorite and chloritoid) schist forms the lowest unit. Outcrops typically weather rusty-brown or black. Bedding is indistinct, but may be marked by discontinuous lenses of quartz that constitute 5 to 10 percent of the rock. A fresh specimen has a somewhat greasy appearance; locally the rock is conspicuously graphitic. In places (e.g. STOP 4) a few feet of quartz-pebble conglomerate are present at the contact with the Precambrian gneiss. The unit is nowhere more than 150 feet thick.

A brown-weathered quartz-albite-muscovite-biotite-microcline granular schist overlies the garnet schist. Beds range from a few inches to several feet in thickness. The contact with the garnet schist is sharp. The upper unit in the Hoosac has a maximum thickness of 150 feet. Both units thin southward and are not present south of Route 116 (Savoy Road).

Dalton Formation

In the northwestern part of the quadrangle the Hoosac Formation is stratigraphically overlain (but structurally underlain) by the Dalton Formation. The Dalton is typically a fissile to flaggy quartz-microcline-muscovite-biotite-albite granular schist or gneiss. Beds are conspicuous, of consistent thickness, and are marked by mica-rich partings. The rock weathers light tan to brown. Clean quartzites, much like those of the Cheshire Quartzite, are locally interbedded with the feldspathic Dalton rocks. These clean quartzites range in thickness from a few inches to as much as 20 feet. In the vicinity of STOP 3, the Dalton and the upper unit of the Hoosac appear to be interbedded and grade laterally into one another. However, to the northwest the Dalton is in sharp contact with the older Hoosac and there is no interbedding of the two lithologies.

South of Route 116, the Dalton lies with angular unconformity on the Precambrian rocks. Here, in addition to the granular schist and gneiss, the Dalton also contains a basal quartz-pebble conglomerate. Both the grain size and the thickness of the conglomerate increase to the south.

The maximum thickness of the Dalton is probably about 500 feet in the Windsor quadrangle; it thins eastward, having thicknesses of 20 feet at STOP 3, and about 2 feet half a mile east of STOP 3.

Cheshire Quartzite

The Cheshire Quartzite is a massive white, pink-weathered or buff-weathered quartzite. Generally it is at least 99 percent quartz. Bedding is recognized only locally where it is faintly indicated by the presence of mica-rich partings. The maximum thickness of the Cheshire is about 100 feet; it thins eastward. At STOP 3 it is about 30 feet thick and where last seen in the eastern extension of the Savoy Hollow Brook syncline (Figure 1), it is 2 feet thick.

Shelburne Marble

The Shelburne Marble is a medium-grained, white-weathered calcite marble. Beds range from 1 inch to more than a foot in thickness, are pronounced, and are marked by thin layers rich in muscovite. Calcite constitutes as much as 99 percent of the rock. Accessory minerals concentrated along bedding planes include quartz, albite, and muscovite. The Shelburne is estimated to be less than 100 feet thick in the Windsor quadrangle.

Berkshire Schist

The Berkshire Schist lies unconformably on all the older rocks of the western sequence in the quadrangle. At several localities within the quadrangle, the Berkshire rests directly on Precambrian gneiss. The predominant lithology is either a quartz-albite-muscovite-biotite-garnet granular schist (STOP 2) or a muscovite-quartz-garnet-chlorite phyllite. Chloritoid and paragonite may be locally abundant in the more aluminous phyllite. All of the rocks of the formation are gray to black because of abundant graphite.

Locally predominant lithologies include sulfidic rusty-weathered quartz-albite-muscovite-biotite-sulfide mineral schist (STOP 4), graphitic quartzite (STOP 4), and black graphitic marble which contains appreciable quartz, albite, and muscovite. Bedding is conspicuous in all rocks except the phyllite. Berkshire rocks are distinguished from the mineralogically similar upper unit of the Hoosac (western sequence) by the presence of more distinct bedding and by a generally rustier appearance on weathered surfaces.

Granitic Gneiss

Granitic gneiss constitutes the lowest unit of the eastern Paleozoic sequence. This unit is apparently in fault contact with Precambrian rocks in the northern part of the quadrangle; this contact has not been observed in either the northern or southern part of the map area. However, the writer believes this body of rock in the Windsor quadrangle and the southern part of the North Adams quadrangle (Herz, 1961) has been incorrectly correlated with the Stamford Granite Gneiss at Stamford, Vermont. There the Stamford is unconformably overlain by the Dalton Formation and is clearly Precambrian.

The granitic gneiss in the Windsor area is a microcline-quartz-biotite-plagioclase augen gneiss. Garnet and muscovite are rare accessory minerals but are more common in the southern part of the quadrangle (STOP 6). The augen texture is locally obscured by intense crushing, the feldspar augen being drawn out to several inches in length. Uncrushed augen commonly are as much as one inch long. The contact with the overlying rocks is sharp; interbedding of the granitic gneiss and the basal rocks of the Hoosac Formation (eastern sequence) is everywhere restricted to an interval of only a few feet.

Hoosac Formation (eastern sequence)

The granitic gneiss is overlain on the east by a sequence of rocks assigned to the Hoosac Formation. In the northeastern part of the quadrangle, the granitic gneiss is overlain by a few tens of feet of quartz-pebble and polymictic-pebble conglomerate. The conglomerate grades upward by transition and interbedding to quartz-albite-muscovite-biotite-chlorite medium-grained schist. In the southeastern part of the quadrangle, the granitic gneiss is overlain by quartz-microcline-albite-biotite gneisses with accessory epidote, muscovite, magnetite, and garnet. Locally, quartz-feldspar quartzites predominate.

The basal gneisses are overlain at STOP 7 by quartz-albite-muscovite-biotite granular schist, typical of much of the Hoosac Formation. This schist unit pinches out northward. Above the albite schist is a thin (200 feet) unit of coarse-grained quartz-muscovite-paragonite-garnet-chlorite-chloritoid schist (STOP 8). This lithology is correlated with the basal garnet schist of the Hoosac in the western sequence although it is slightly more aluminous.

The garnet schist of the eastern sequence is overlain by 4,000 feet of quartz-albite-muscovite-biotite(or chlorite, or biotite and chlorite) schist that is typical of the Hoosac Formation in Massachusetts.

ROAD LOG AND STOP DESCRIPTIONS

Assemble at 8:30 A.M. in the parking lot of the Adams Market, diagonally opposite the U. S. Post Office at the intersection of Routes 8 and 116 at the south end of Adams, Massachusetts.

NOTE: Coordinates for stops are based on the Massachusetts coordinate system.

Mileage

0.0 Proceed east up hill on Route 116 (Savoy Road).

0.4 Small outcrops on left of Clarendon Springs Dolomite in inverted position.

0.9 The house across the gully to the right (west) is built on a large kame terrace.

1.2 STOP 1 (58.93N - 16.44E) Enter the woods on the east side of the road about 100 feet north of the hydrant. The outcrop is about 200 feet east of the road.

Here is exposed a series of ledges of well-bedded, fine- to medium-grained calcite marble of the Shelburne Marble. Beds range from 1/10 inch to 2 feet in thickness. In the lower part of the outcrop is a large open fold with an east-over-west shear sense. An earlier lineation is present and wraps around the axis of this later fold that refolds isoclinal folds that are related to the formation of the Hoosac nappe. The axial surfaces of the early folds are nearly everywhere parallel to the bedding. They strike north and dip 20 to 30° east. The axes plunge gently to the north. Locally this attitude is slightly disturbed by the later folding.

Return to cars and continue on Route 116.

1.4 STOP 2 (58.75N - 16.44E) Roadcut on right (southwest) side of Route 116.

Here is exposed gray quartz-albite-muscovite-biotite graphitic schist of the Berkshire Schist. At the north end of the outcrop are recumbent isoclinal folds related to the Hoosac nappe. Schistosity parallels the axial surface of these folds. Twenty feet south of the prominant isoclinal folds are more open folds that fold schistosity. The later folds in the Berkshire are normally chevron in style. Both the bedding and the schistosity project beneath the rocks at STOP 1. Here the Berkshire is interpreted to lie with stratigraphic unconformity on the Shelburne Marble, both formations being upside down in the overturned limb of the Hoosac nappe. Thus, in this area, the unconformity beneath the Berkshire has cut down through the Bascom Formation. The contact between the Berkshire and Shelburne is not exposed. One half mile to the west, the Bascom and Berkshire are in contact

Continue on Route 116.

1.7 Cheshire town line.

3.2 The washed bank to the left (north) of the road consists of brecciated Cheshire Quartzite with a ferruginous cement. This breccia zone marks the trace of the Hoosic thrust (Herz, 1961; Norton, 1967).

3.4 Roadcut on left (north) side of road of Berkshire Schist slightly above the Hoosic thrust and below the Savoy Hollow Brook thrust (Figures 1 and 3).

3.6 Start up windy road with brook along left (north) side of the road. Berkshire Schist is exposed in the lower part of the brook.

3.8 Berkshire Schist in brook below Savoy Hollow Brook thrust is overlain by Cheshire Quartzite above thrust. The Cheshire is in turn structurally overlain by the Dalton Formation in inverted sequence.

3.9 Outcrops of Cheshire may be seen on the north bank of the brook.

4.2 Long roadcut of Berkshire Schist. We are below the Savoy Hollow Brook thrust at this point.

4.3 Savoy town line. Still in Berkshire Schist.

4.8 Scattered outcrops on both sides of the road of Precambrian gneisses. Trace of Savoy Hollow Brook thrust is about 100 feet north (left) of the road.

5.2 Outcrop on left side of road of flaggy Dalton feldspathic gneiss.

5.4 STOP 3 (58.00N - 17.91E) Large roadcut at the height of land between Adams and Savoy.

Exposed on the road is pink, white, and buff Cheshire Quartzite. This outcrop, along with other similar exposures in the town of Cheshire, was originally a source of glass sand. The west end of the outcrop consists of friable quartzite whereas the eastern end is tough vitreous quartzite. The cause of this difference is not known. Neither tectonic crushing nor the presence of accessory minerals appear to explain the relative cohesion of the rock. Although the Cheshire is generally unbedded, bedding may be recognized with difficulty at the west end of the outcrop on the south side of the road. One hundred and fifty feet east of this exposure, on the north side of the road, is an outcrop of rock that most closely resembles the massive albite schist of the Hoosac Formation of the western sequence. Less than 1,000 feet to the west this rock type is interbedded with and passes laterally into flaggy rocks typical of the Dalton Formation. About 100 feet

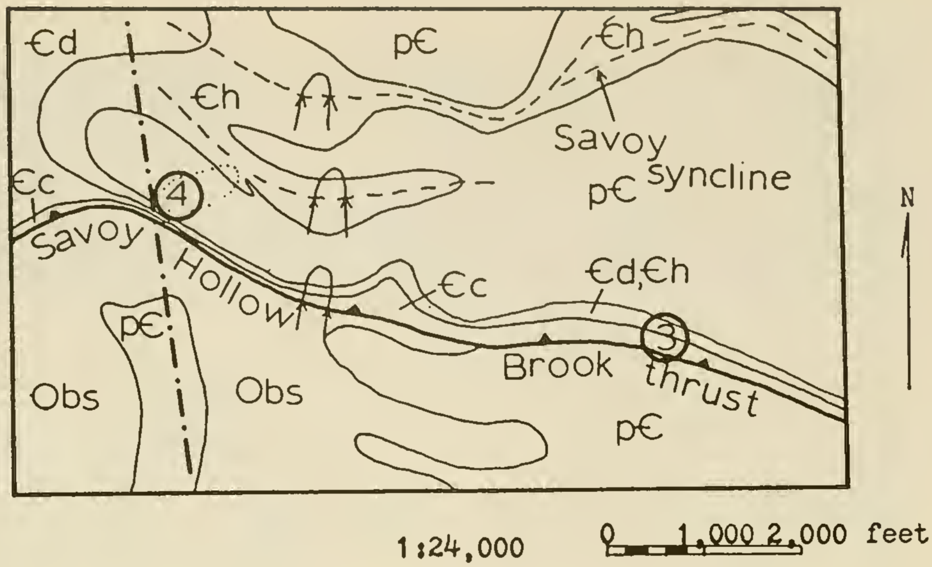

Figure 3. Generalized geologic map in the vicinity of STOPS 3 and 4. Route 116 is nearly coincident with the thrust fault. Dashed and dotted line is the town line between Savoy (east) and Cheshire (west), Massachusetts. Symbols are the same as those used for Figure 1.

north of this outcrop of Hoosac is an outcrop of well banded Precambrian gneiss, similar to that seen at the next stop. In this area the Precambrian-Cambrian contact unfortunately occurs in a swale and can not be seen. South of the road, the Cheshire is apparently in fault contact (the Savoy Hollow Brook thrust) with the Precambrian gneisses. This same inverted sequence may be followed southeastward for 8,000 feet.

Continue on Route 116.

5.5 The outcrop on the right (south) side of the road may be a sliver of Berkshire beneath the thrust. These rocks resemble those seen at STOP 2.

5.6 Large driveway. Turn cars around and head west on Route 116.

5.8 Pass by STOP 3.

6.9 STOP 4 (58.12N - 17.37E) Just before Cheshire town line and roadcut in the Berkshire Schist. We shall look at the roadcut and then traverse up the steep hill to the north.

A variety of rocks within the Berkshire Schist are exposed on the road. They include quartzites, quartz-albite-muscovite-biotite schist, and quartz-muscovite schist. All the rocks are graphitic and contain abundant sulfide minerals, predominantly pyrite and pyrrhotite, which cause rusty yellowish-brown to black weathering.

The internal structure of the outcrop is complex and has defied rigorous interpretation. Although folds of similar style in similar rocks range in attitude fro horizontal to vertical, no system of later folds is evident to explain these differences in attitude. In places, minor faulting has disrupted the beds. The complex structure is probably explained by the presence of the Savoy Hollow Brook thrust which trends up this valley parallel to the road. The trace is about 50 feet north of the outcrop and the fault surface projects just over your head at this point.

From the northeast end of the outcrop, traverse across the field on a bearing of N.20°E. a distance of about 375 feet to the break in slope and a large cliff. The traverse crosses unexposed Dalton and Cheshire in inverted position. At the cliffs are well-banded Precambrian quartz-feldspar-biotite gneisses. The general absence of muscovite in these Precambrian gneisses distinguishes them from the Dalton and Hoosac gneisses with which they may be confused. Minor, gray-weathered, somewhat feldspathic quartzites and massive feldspar-quartz-biotite gneissic granulites will be seen on the traverse up the hill. At the base of the cliff, two styles of folding may be seen:

1. Isoclinal recumbent folds with a gentle northerly plunge. These are presumably related to the formation of the Hoosac nappe. Schistosity is axial planar to these folds.

2. More open folds of small amplitude with a consistent east-over-west shear sense fold the schistosity. The axial planes of these later folds strike northerly and dip about 30° east; the axes plunge gently to the north.

These two fold generations correspond to the two sets of folds seen at STOP 1. The later folds are nearly coaxial with the isoclinal folds and appear also to be related to the nappe but they must have formed after the emplacement of the nappe because their shear sense is consistently east-over-west, regardless of position with respect to the larger recumbent syncline we shall see at the top of the traverse (Figure 3). In places, the compositional layering can not be followed for more than a few tens of feet. The gneissosity in those rocks which are generally more micaceous is interpreted to have formed from the transposition of earlier compositional banding during the formation of the nappe.

Continue up the cliffs of Precambrian gneiss until a break in slope is reached with a broad plateau. On the southwesternmost corner of the plateau is a low 30 feet long outcrop which exposes quartz-pebble conglomerate resting unconformably on the Precambrian gneisses. The contact may be traced from outcrop to outcrop and has itself been folded in a left handed sense with east-over-west transport. At this outcrop, the contact trends N.25°W. and dips 33°NE. The stretched pebbles have their maximum dimension oriented about N.15°E. which is essentially parallel to the axes of both the early recumbent isoclinal folds and the later open, east-over-west shear folds. The rocks exposed to the northeast and north are rusty-brown to black weathered quartz-muscovite-albite-garnet-biotite schist typical of the basal unit of the Hoosac of the western sequence.

If time permits, we will walk northeast to the next break in slope where the Precambrian gneisses rest on the Hoosac in inverted sequence. Thus we are on the normal limb of a southwest-facing recumbent syncline. This syncline closes about 2,000 feet to the east.

Return to cars. Continue west on Route 116.

7.7 Left on Fales Road. Poorly marked turn - be careful.

7.8 Small outcrop on left of Berkshire Schist. The hill 4,000 feet to the west is all Berkshire Schist.

8.7 Left on Sand Mill Road.

8.8 Bear left on dirt road.

8.9 Cross trace of Hoosic thrust onto upper plate. Breccia of Cheshire Quartzite is exposed in brook to left.

9.2 Join Windsor Road. Gorge on left in Cheshire Quartzite. Cross Dry Brook - so named because in the lower reaches, this brook flows over Bascom Formation carbonate rocks and solution channels cause a loss of surface water to subterranean channels.

9.4 Small unnamed brook on left contains an outcrop of Berkshire Schist occupying the core of Dry Brook syncline (Figures 1 and 4), an isoclinal recumbent fold, overturned to the southwest. We will see rocks higher in the brook at STOP 5A.

9.6 STOP 5 (57.11N - 17.14E) At the break in the fence line. This stop will involve a 5,000 feet long traverse. Bring your lunches. We will eat at the outcrops with a fine view of the Hoosic Valley and Mount Greylock. From the break in the fence, proceed northerly across the small field to an old woods road and follow it uphill. About 600 feet from Windsor Road, branch left on an obscure path and continue another 150 feet to a brook. Proceed up the brook. About 320 feet upstream are slumped outcrops of Dalton. Five hundred feet upstream the stream bed steepens through a jumble of boulders, most of which are Dalton. Directly above the boulders is a large outcrop of Precambrian gneiss (STOP 5A), the base of which is highly sheared and marks the position of the inverted unconformity. Note the angularity of the contact. About 100 feet northwest of this point, the contact is clearly exposed in a slumped outcrop. Note the quartz pebbles as much as 6 inches in diameter in the loose boulders.

Walk north about 750 feet across small scattered outcrops of Precambrian gneisses to the ridge on the high meadows. You may follow the traverse on Figure 4.

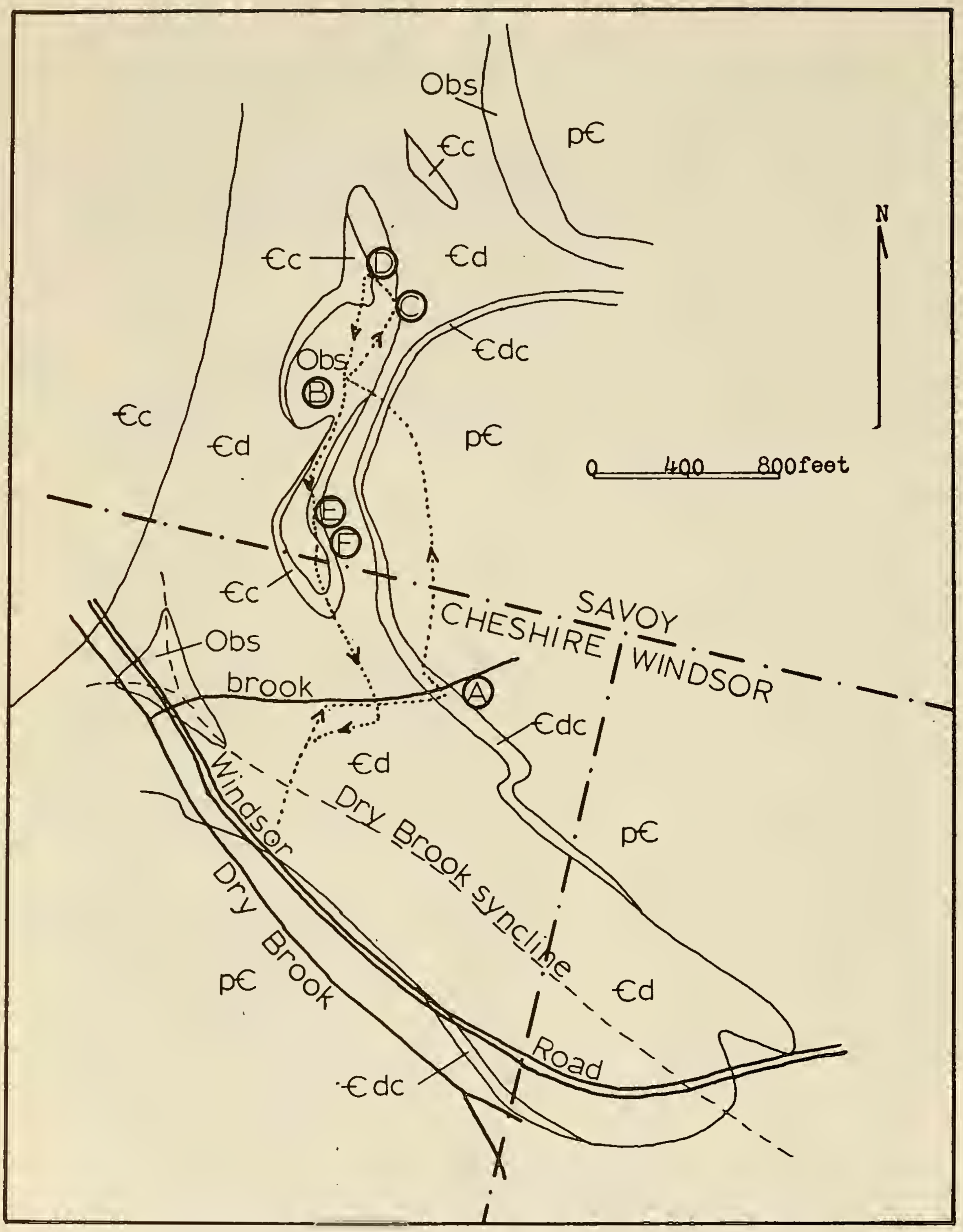

Figure 4. Geologic map in the vicinity of the Dry Brook syncline, Windsor quadrangle, Massachusetts. Circled letters indicate outcrops of particular interest discussed in the text under STOP 5. Symbols are the same as those used for Figure 1.

LUNCH

Points of Interest

1. The valley to the southwest is occupied by the recumbent Dry Brook syncline, overturned to the southwest. Our traverse has brought us through the inverted limb.

2. Mount Greylock, composed of allochthonous Taconic rocks, may be seen across the Hoosic Valley on a bearing of N.35°W.

3. Stafford Hill, composed of Berkshire Schist, is the nearest hill on a N.65°W. bearing. The valley beyond Stafford Hill is occupied by the western carbonate sequence.

4. The bench 300 feet to the northwest is occupied by an open syncline, overturned to the west. See Figure 4 for the plan of our traverse.

STOP 5B Center of syncline. Berkshire Schist occupies the core of this syncline throughout its entire length. Here the rock consists of brown-weathered quartz-albite-muscovite-biotite-tourmaline schist.

STOP 5C Here Berkshire Schist is in contact with medium-grained quartzites of the Dalton Formation. Some of the beds contain quartz pebbles as large as $\frac{1}{2}$ inch. This locality is on the overturned limb of the syncline.

STOP 5D This outcrop, 175 feet northwest of STOP 5C, is on the normal limb of the syncline. Here, Berkshire is in contact with the Cheshire Quartzite.

STOP 5E S.15°W. 450 feet from STOP 5B, the contact between the Cheshire and the Dalton is clearly exposed. At this point we are on the overturned (eastern) limb of the fold. The Dalton is somewhat atypical here in that it is non-rusty weathered. The outcrop about 125 feet northwest of this point is Dalton. Thus either the Berkshire is very thin through this part of the fold, or it is not present at all.

STOP 5F Two hundred and fifty feet S.5°W. is the contact between the Berkshire and the Cheshire on the nearly (at this point) flat lying, overturned limb of the fold.

Proceed S.25°E. following a series of outcrops of Berkshire which occupy the core of the syncline. The hinge line beneath the Berkshire emerges before the brook is reached (about 650 feet from STOP 5F) and projects into the air over the brook. Cross the brook and continue about 100 feet to old woods road and proceed down to cars.

Continue on Windsor Road.

9.8 Windsor town line. Road becomes Cheshire Road.

10.0 Small outcrop of Precambrian gneiss on left.

10.2 Large outcrop of Precambrian gneiss on left, 100 feet off road.

10.4 Home of Al Chamberlain on right, handcrafted from the ubiquitous glacial boulders of Cheshire Quartzite.

12.3 Intersection with paved Savoy Road (Route 8a). Turn right (south).

14.1 Intersection with Route 9 at the center of Windsor, Massachusetts. Turn left (east) on Route 9.

14.6 Whaleback outcrop off road to right exposes granitic gneiss of the eastern sequence. Microcline augen in this outcrop are as large as $2\frac{1}{2}$ inches in length.

14.8 Cross Savoy Hollow Road (on left)-Humes Road (on right).

15.0 STOP 6 (55.35N - 18.35E) Pull cars to right. Roadcut on left (north) side of road.

This outcrop is located within what is mapped as granitic gneiss. It consists of a microcline-quartz-biotite-plagioclase gneiss with accessory garnet, muscovite, and magnetite. The augen of microcline which are characteristic of this unit are crushed and drawn out into lenses at this locality. The unit, aside from the texture, is very homogeneous, both perpendicular and parallel to strike. The unit is believed to be correlative with the Bull Hill Member of the Cavendish Formation of Doll and others (1961) which forms the base of the Paleozoic section in several areas in Vermont.

Continue east on Route 9.

15.8 STOP 7 (55.04N - 18.65E). Roadcut on right side of road.

Here, the basal Hoosac gneisses, composed of albite, quartz, muscovite, biotite, and epidote, are in contact with the overlying quartz-albite- and albite-quartz-muscovite-biotite schist of the Hoosac Formation. The schist is slightly graphitic and weathers a characteristic brown or rusty-brown. Bedding in the schist is obscure, but is well-displayed in the light colored gneisses. Both of the rock types present here pinch out northward between the granitic gneiss (STOP 6) and the overlying garnet schist member of the Hoosac Formation (STOP 8) and are not present at the north edge of the quadrangle.

The pronounced schistosity in the schist is axial plane to isoclinal steeply plunging folds that are best seen on the top of the weathered outcrop. The numerous quartz lenses that outline these folds are interpreted as quartz-rich lenses in the original sediment. Late southwest plunging open reverse drag folds may be seen in the fresh roadcut surface. These folds fold schistosity and only rarely display a weak axial plane cleavage.

Continue east on Route 9.

16.1 Reverse direction on Route 9.

16.5 Pass by STOP 7.

17.5 Turn right (north) on Savoy Hollow Road (dirt).

17.7 Turn right (east) on Shaw Road.

17.8 Whaleback outcrop in road of granitic gneiss.

17.9 Whaleback outcrop in road of granitic gneiss.

18.4 STOP 8 (55.37N - 18.63E) Top of rise in road. Outcrop extends up ridge to north of road.

This exposure is in the garnet schist member of the Hoosac Formation and directly overlies the albite schist seen at STOP 7, an outcrop of which may be seen 100 feet west of STOP 8. The rock is a coarse-grained quartz-muscovite-paragonite-garnet-chlorite-chloritoid schist. Minor amounts of albite may be present in some beds. The rock has a very strong schistosity which is folded unsystematically by folds that only locally display a weak slip cleavage. The rock here is correlated with the basal garnet schist of the Hoosac of the western sequence (STOP 4). To the north, it is tentatively correlated with the garnet-mica schist member of the Cavendish Formation of Doll and others (1961) in the Rowe, Massachusetts quadrangle (Chidester and others, 1967), with the Heartwellville Schist of Skehan (1961) in southern Vermont, and with the Gassetts Schist Member of the Cavendish Formation in central Vermont (Doll and others, 1961).

END OF TRIP

To rejoin Route 9, continue on dirt road to

19.4 Outcrop of typical albite schist of the Hoosac Formation.

20.1 Intersection of High Street (formerly Shaw Road) and High Street Hill Road. Turn right. Outcrop at the corner is the basal graphitic schist of the Rowe Schist (Figure 2).

20.2 High tension lines.

20.6 Rejoin Route 9. For Connecticut, eastern Massachusetts, New Hampshire, and Maine, turn left and proceed to Northampton and Interstate 91. For Vermont, turn right and return to Windsor. Take Route 8a north. For New York, turn right and follow Route 9 to Pittsfield.

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